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# RECORDS OF PLEISTOCENE GLACIERS IN THE MEDICINE BOW AND PARK RANGES[1]

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ABSTRACT

The glacial features of the Medicine Bow and Park ranges of Wyoming and Colorado record two stages of Pleistocene glaciation. In both areas well-developed Wisconsin moraines have been mapped. Records of pre-Wisconsin ice have been found in the form of old drift deposits located above and beyond the reach of Wisconsin glaciers. A restoration of the late Pleistocene landscape is shown in the color plate accompanying this report.

In addition to the record of two glacial stages found in both mountain regions, the Park Range presents an interesting contrast between east- and west-slope glaciers. The tongues of ice which descended the west side of the range were longer than those on the east and terminated approximately 1,000 feet lower. These differences are due in part to favorable exposure to moisture-bearing westerly winds.

West of the Front Range of the Rocky Mountains, and near the Wyoming-Colorado line, are two high mountain areas known as the Medicine Bow and Park ranges (Figs. 1, 5, and 10). Located between two main transcontinental highways, these snow-clad mountains have been little visited by the thousands of travelers racing across the country. During the past few years, however, an excellent mountain road has been completed across the Medicine Bows from Laramie to Saratoga and this has attracted an increasing number of visitors. No road crosses the Park Range except at the south

[1] The author wishes to express appreciation to Clark University for financial assistance received through grants from the William Libbey Fund in 1933 and 1936.

through Rabbit Ears Pass. In order to reach the high mountain country it is necessary to follow the foot and horse trails.

The author visited both ranges during the summers of 1932, 1933, and 1936 for the purpose of mapping the glacial features and studying the erosional history of the North Platte River and its tributaries. The present paper deals with glacial phenomena which have proved to be extremely interesting in themselves and valuable aids

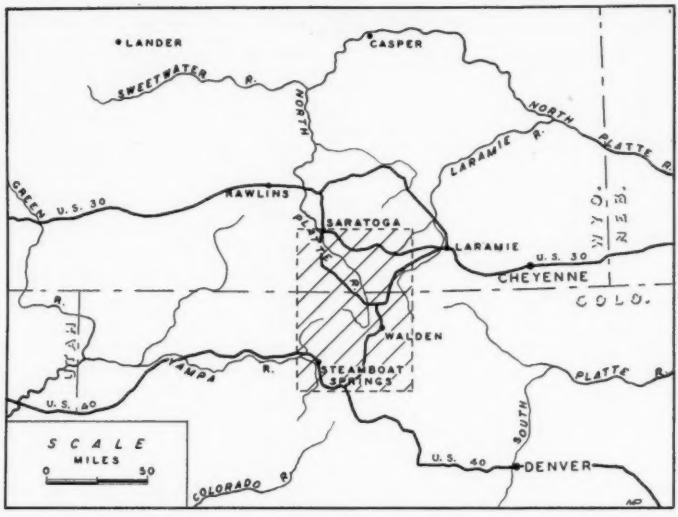


FIG. 1.—Location of the region of study. The main highways and important drainage lines are indicated. The shaded portion represents the area covered by Plate I.

in the interpretation of the physical history of the Rocky Mountain region, now in progress in association with Wallace W. Atwood, Sr.

## MEDICINE BOW RANGE

Like so many ranges of the Rocky Mountains, the Medicine Bow was severely glaciated during the Pleistocene epoch. Only the higher elevations rose above the ice fields which formed on the mountain slopes. Every major valley was occupied by glaciers and every master-stream carried an abundance of débris far out onto the surrounding plains. Many of the picturesque features of the Medicine

PLATE I

A RESTORATION OF THE LATE PLEISTOCENE LANDSCAPE
IN THE MEDICINE BOW AND PARK RANGES

Glacial geology by Wallace W. Atwood, Jr.
Painting by Eugene Kingman

Bow peaks, hereafter referred to as the Snowy Range, were produced at this time. In an attempt to picture the late Pleistocene landscape, a restoration of Wisconsin glaciers has been prepared to accompany this report (Pl. I). It is hoped that this painting by Eugene Kingman, based upon field evidence gathered by the author, will help the reader to visualize the conditions in these ranges during the presence of the ice.

### WISCONSIN GLACIATION

*Libby Glacier.*—To the east of Medicine Bow Peak[2] is a large, plateau-like area 10,500–11,000 feet above sea-level. During the Wisconsin stage this region served as the catchment area for Libby Glacier, which at the time of maximum advance was over 9 miles in length and covered approximately 30 square miles. The scouring effect of the ice is well exhibited throughout this catchment area and eastward along Libby Creek. Many rock-basin lakes lie partially hidden between highly polished rock surfaces, giving the region the appearance of having been only recently deprived of glacial ice (Fig. 2).

Although the Libby Glacier left little morainic material on the plateau country, it deposited large amounts of heavy drift in the valley of Nash Fork and Libby Creek. The main highway from Centennial to University Camp crosses several of these glacial deposits. A short distance east of Libby Lodge the road cuts through a perfect morainic ridge made of boulder till containing well-striated stones. Another good example of moraine is situated a few hundred yards northeast of the University of Wyoming summer camp. Here the hummocky topography and undrained depressions of frontal deposits are well developed. Striae are common on the fine-grained, dark-colored rocks. A somewhat smaller area, but one displaying equally fine morainic characteristics, occurs 3 miles southeast of Medicine Bow Peak. All these areas are indicated on the map of glacial features (Fig. 4).

While it is quite impossible to determine the exact upper limit of ice action on the slopes of the Snowy Range, due to subsequent

[2] All places referred to in the description of Medicine Bow glaciers may be located on the index map (Fig. 5).

erosion and the accumulation of talus material, it is probably safe to assume that ice stood as high as 11,500 feet. Above this elevation the mountain slopes do not exhibit oversteepened walls due to ice-plucking. The lower limit of the ice is more easily determined by the frontal moraines which in Libby Creek are found at approximately 8,500 feet (Fig. 3). Thus, it may be seen that the ice of Libby Glacier had a vertical extent of 3,000 feet. If the surface gradient

FIG. 2.—The catchment area for Libby and French Creek glaciers on the east side of the Snowy Range. The photograph was taken from the summit of Medicine Bow Peak looking southward across Mirror Lake to the high plateau which extends to the Wyoming-Colorado line.

of the glacier were calculated on the basis of these figures, it would be approximately 315 feet per mile. In the case of Libby Glacier, however, the position of lateral moraines offers a more satisfactory basis for such estimates. Well-developed marginal ridges occur at 10,500 feet, 9,500 feet, and again at 8,500 feet elevation—all within a distance of $7\frac{1}{2}$ miles. By using these figures we find the surface gradient to be about 265 feet to the mile.

To the east of the frontal moraine of Libby Glacier extensive outwash deposits occur. These consist chiefly of gravel and cobblestone material derived from quartzites of the Snowy Range. The highway leading eastward from Centennial follows a portion of the out-

wash plain which slopes to the southeast, at approximately 80 feet to the mile. In places recent stream alluvium covers the gravels and elsewhere wash from neighboring mountains blends with the glacial material, making it difficult to determine the exact extent of the outwash.

*French Creek Glacier.*—Although deriving most of its ice from the same catchment area as Libby Glacier, the French Creek Glacier

FIG. 3.—The valley of Libby Creek northwest of Centennial. The white line marks the margin of Libby Glacier. Pre-Wisconsin glacial deposits are located above and beyond the limits of Wisconsin ice.

advanced in almost the opposite direction (Fig. 4). Confronted by a plateau barrier to the southeast, the ice was forced to follow the deep V-shaped canyon of South French Creek. From the highway which traverses the area south of the Snowy Range, numerous striated surfaces are visible. The direction of these markings is invariably to the southeast. Upon reaching the main valley, however, the ice changed direction and pushed southwesterly for a distance of more than 8 miles.

Morainic accumulation in South French Creek is limited to a thin mantle over most of the valley bottom and a few patches of heavy drift which mark frontal positions of the ice. At the point of separation of the Libby and French glaciers there is a small area of hum-

mocky topography on the plateau barrier at 10,500 feet (Fig. 6). This morainic material is 2,200 feet above the frontal deposits which appear 9 miles down the South French Creek valley. From these figures the surface gradient of the French Creek Glacier is estimated

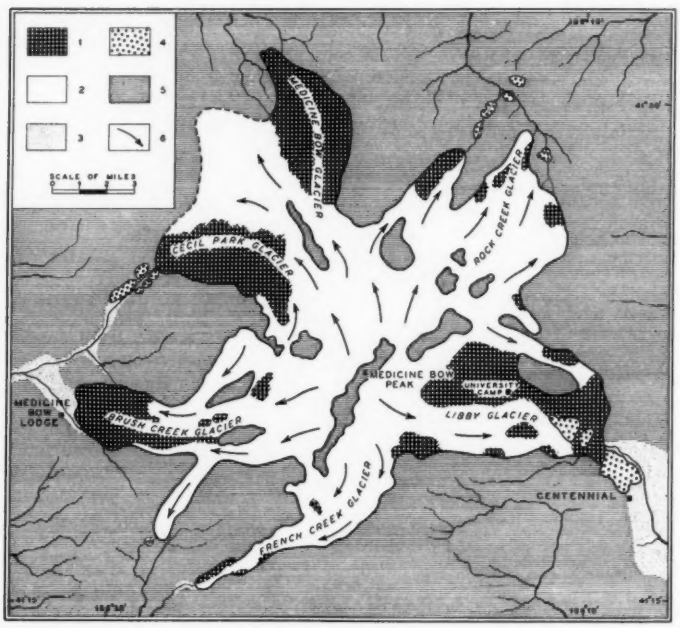


Fig. 4.—Glacial features of the Medicine Bow Range: (1) heavy Wisconsin moraine; (2) areas covered by Wisconsin glaciers; (3) Wisconsin outwash deposits; (4) pre-Wisconsin moraine; (5) areas unglaciated during Wisconsin time; and (6) direction of ice movement.

to be approximately 245 feet per mile. The even contour of the valley floor suggests that the gradient was uniform.

Owing to the narrow V-shaped character of South French Creek valley below the terminal position of the glacier, very little outwash remains. Two small terrace remnants were recognized, but these do not continue far down the valley. If French Creek Glacier supplied much outwash material, it has since been removed.

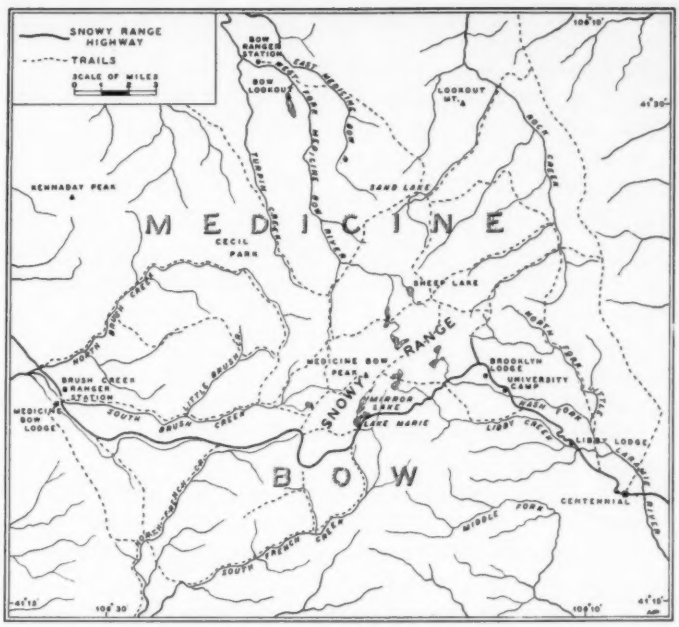


FIG. 5.—Index map of the Medicine Bow Range. All places referred to in the description of Medicine Bow glaciers are located on this map.

FIG. 6.—Wisconsin moraine 3 miles southeast of Medicine Bow Peak. Boulder-strewn hills and undrained depressions are abundant in this youthful glacial landscape. A portion of the Snowy Range is visible at the right of the picture.

*Brush Creek Glacier.*—The glacier which occupied the valley of Brush Creek on the western slope of the range had a maximum length of approximately 10 miles. The ice came from a compound catchment area including the headwaters of Little Brush, South Brush, and North French creeks. Ice from these three sources joined about 5 miles east of Medicine Bow Lodge, producing Brush Creek Glacier, which spread to a width of 2 miles in the broader valley to the west, and soon reached its terminal position, less than a mile east of the Lodge (Fig. 4).

The amount of drift remaining in the area formerly covered by the Brush Creek Glacier is much greater than in Libby or South French Creek. Unusually heavy moraines mark the frontal position near Medicine Bow Lodge and large lateral accumulations appear to the east along the highway. Smaller patches occur in upper Brush Creek and in the catchment areas. In the higher country, however, the morainic mantle is rather thin, due to the scouring effect of the ice. The surface gradient of this glacier is estimated to have been approximately 260 feet per mile.

Beyond the terminal moraine outwash begins. The entire valley north and northwest of Medicine Bow Lodge is filled with Wisconsin outwash material. The Brush Creek Ranger Station and the Lodge are both located on these coarse gravel deposits. Recent alluvium frequently mantles the coarse material, thereby making the land more suitable for agricultural use.

*Cecil Park and Medicine Bow glaciers.*—The large catchment area to the northwest of the unglaciated crest of the Snowy Range supplied ice to the Cecil Park and Medicine Bow glaciers (Fig. 4). In contrast to the precipitous cliffs of the eastern side of the mountains, the western slope is fairly gradual with no pronounced cirque walls. Ice filled the valleys and spread over much of the high plateau country to the north of Medicine Bow Peak. One tongue of ice followed the valley of North Brush Creek and terminated west of Cecil Park, from which it receives its name.

Another slightly larger tongue followed the West Fork of Medicine Bow River. The two glaciers were separated from one another by a high ridge which rose as a nunatak above the ice. To the northwest of this obstruction the glaciers joined and continued as one for some distance. Heavy down timber, swamps, and the absence of

trails made it difficult to determine the exact extent of the ice in this direction. In consequence a dotted line on the map indicates the approximate frontal position in this locality (Fig. 4).

Thick till, containing well-striated stones, characterizes the marginal moraines of both the Cecil Park and the Medicine Bow glaciers. In the vicinity of Bow Lookout and Bow Ranger Station excellent examples of hummocky topography were observed. The large amount of morainic débris is indicative of the severity of ice work in the high country where large areas of smoothed and polished bed rock occur.

In contrast with Libby, French, and Brush Creek glaciers, the surfaces of Cecil Park and Medicine Bow glaciers had gradients of 150–200 feet per mile. This gentle slope is undoubtedly due to the fact that the ice spread out on a plateau-like surface instead of following restricted valleys as was the case with the other glaciers.

Outwash from Cecil Park Glacier followed the valley of North Brush Creek to merge with the wash from Brush Creek Glacier 2 miles north of Medicine Bow Lodge. There was no great amount of material deposited at this locality and consequently only a narrow strip remains today. The outwash from Medicine Bow Glacier followed the east and west forks of the Medicine Bow River.

*Rock Creek Glacier.*—With ice acquired from the catchment area of the Medicine Bow Glacier and the high country to the northeast of the Snowy Range, Rock Creek Glacier moved down the western slope of Rock Creek valley to reach its frontal position approximately 2 miles southeast of Lookout Mountain. In a distance of less than 6 miles, the total length of the glacier, the ice descended from 11,000 feet at its source to 8,500 feet at its terminus. The steep gradient of over 400 feet per mile, while far greater than that recorded by neighboring glaciers, is easily explained by the topography. The high plateau which surrounds the Snowy Range is very narrow at the northeast and consequently glacier ice reached the low country without first having to cross a wide expanse of high land.

## PRE-WISCONSIN GLACIATION

Evidence of pre-Wisconsin glaciation in the Medicine Bow Range is best displayed in the valley of Libby Creek. Here the old glacial deposits are well preserved and easily reached. Other exposures

occur in North Brush Creek beyond the frontal position of Cecil Park Glacier and in the valley of Rock Creek east of Lookout Mountain. None of these occurrences, however, is as satisfactory as the Libby Creek locality for demonstrating the existence of two glacial stages.

The largest area of old moraine is traversed by the highway north and west of Centennial. The mature, gently rolling topography of this ancient drift stands out in striking contrast to the young Wisconsin moraines immediately to the west (Fig. 3). There are no lakes and swamps such as characterize young moraines, and the rocks are more severely weathered than those of the Wisconsin drift a few hundred feet away. A careful search among the heterogeneous materials of the old glacial deposits produced several striated stones. Many of the cobbles were subangular, and some of the larger boulders appeared to have been glacially polished.

Upstream from this largest area of pre-Wisconsin moraine are two smaller remnants, both situated on the divide between Libby Creek and Nash Fork. Again the mature topography of an old moraine attracts attention and subangular striated stones give positive proof of glacial origin (Fig. 7). Located above the reach of Wisconsin ice, these two areas of old drift remain as isolated remnants of an earlier glaciation. While there are insufficient records to reconstruct the early glaciers, it is quite evident that they were slightly larger than those which followed them in Wisconsin time.

### PARK RANGE

The Park Range of Colorado is approximately 45 miles in length and the crest-line elevation varies between 10,000 and 12,000 feet. This north-south range has long served as an effective barrier to moisture-bearing winds and at times has been heavily mantled with perennial snowfields in which ice formed. During the Wisconsin stage of the Pleistocene this high mountain country was almost entirely covered with glaciers (Pl.I). In the central and northern portions of the range only serrate peaks rose above the ice, while in the southern section an ice cap covered the Continental Divide. A score of glaciers descended the mountain valleys to terminate on the lowlands 3,000–4,000 feet below. No permanent ice fields remain in these mountains today.

WISCONSIN GLACIATION

*Elk, South Elk, and Hinman glaciers.*—The largest of the many glaciers which radiated from the Park Range was the one which filled the Elk River valley and its tributaries west of the Continental Divide. Sixteen catchment basins located in the central and western part of the range contributed to this one glacier (Fig. 9). Only the serrate peaks of the Sawtooth Range and the high intervalley areas north and south of the present river rose above the ice.

FIG. 7.—Pre-Wisconsin moraine 3 miles northwest of Centennial. The subdued topography shown in this photograph is characteristic of the old drift wherever observed in the Medicine Bow region. Large boulders and numerous striated stones give indication of the glacial origin.

The chief supply of ice for the Elk Glacier came from the cirques west and southwest of Mount Zirkel.[3] Here, close to the crest of the range, heavy snows fell and ice fields formed. As a result, precipitous cirque walls were produced, the mountain valleys were scoured and polished, rock-basin lakes came into existence, and a magnificent mountain landscape was created (Fig. 8). Today this rugged scenery is not visible from the low country to the east or west and comes as a real surprise to the traveler who ventures into the heart of the range. The unimpressive appearance of the mountains from the North Park

[3] All places referred to in the description of Park Range glaciers may be located on the index map (Fig. 10).

highway is perhaps one of the chief reasons why few people visit the region.

In addition to the ice derived from the Sawtooth Range, a substantial amount came from the South Fork of Elk River which received its supply from the high country a few miles north of Mount Ethel. Another ice stream, with its sources near the western

FIG. 8.—A portion of the Sawtooth Range west of Mount Zirkel. The Elk and Encampment glaciers both received ice from this catchment basin. Beautifully polished and striated surfaces are found throughout this high cirque region.

edge of the range, moved southward down the valley of Hinman Creek to join the main glacier at Hinman Park. At the time of maximum extension the Elk Glacier reached the western end of Hinman Park, where it deposited an excellent moraine just south of the river. Downstream from this point Wisconsin outwash forms low terraces on either side of the modern flood plain.

Well-polished and striated rock surfaces 1,000 feet above Elk River in the narrows east of Hinman Park indicate that the glacier was at least 1,000 feet thick as it passed that point (Fig. 11). Farther upstream the ice filled the valley to a depth of 1,500 feet.

Four miles east of Hinman Park is one of the best-developed moraines in the Elk Valley system. Here the South Elk Glacier merged temporarily with the main ice stream forming a perfect medial moraine 2 miles long and several hundred feet in height. In the valley of North Fork a heavy mantle of moraine covers much of the country. This large accumulation, far from the frontal position of the ice, may possibly be explained by partial stagnation produced when the ice from the north and western slopes of the Sawtooth Range joined Hinman Glacier. Except for these few areas of heavy moraine just mentioned, the entire region traversed by the Elk Glacier and its tributaries is relatively free of morainic débris. Much of the high country is scoured and polished although a few sections are veneered with a thin mantle of drift.

During the retreat of the Elk Glacier the outwash which now covers the floor of Hinman Park was deposited. A few moraine hills rise above the gravel mantle, but the general landscape is that produced by stream alluviation. In postglacial time the Elk River has cut through this wash producing a young valley ranging from 15 to 25 feet in depth. Smaller areas of Wisconsin outwash occur still farther upstream and exhibit much the same characteristics.

*Big Creek Glacier.*—Originating on the western slope of Mount Ethel, Big Creek Glacier moved down the valley for a distance of 10 miles. Throughout most of this course glacial evidence is limited to smoothed and polished rock surfaces. Near the terminal position, however, excellent morainic ridges appear on the valley walls. The traveler who takes the ridge trail from Big Creek to Swamp Park follows one of these moraines for over a mile (Figs. 9 and 10). On the north side of the valley a well-developed lateral forms the divide between Big and Reed creeks. During an earlier stage of glaciation, to be discussed later, ice pushed across this divide and nearly reached Elk Creek below Hinman Park. The frontal deposits of Big Creek Glacier occur in the valley 4 miles southeast of Clark. Beyond these hills of drift, small outwash terraces border a narrow strip of recent alluvium.

*Mad Creek Glacier.*—Like Big Creek, Mad Creek derived some of its ice from cirques on the western slope of Mount Ethel. Additional ice came from the Continental Divide south of the crest-line peaks.

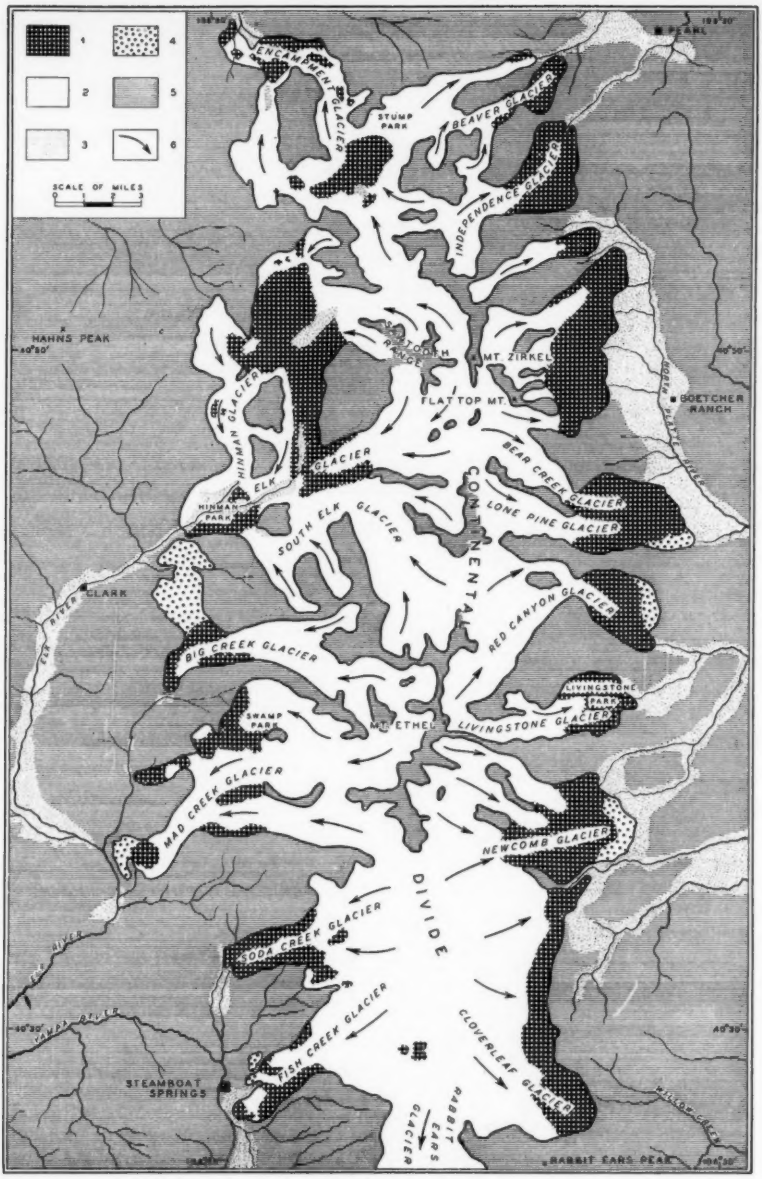


FIG. 9.—Glacial features of the Park Range: (1) heavy Wisconsin moraine; (2) areas once covered by Wisconsin glaciers; (3) Wisconsin outwash deposits; (4) pre-Wisconsin moraine; (5) areas unglaciated during Wisconsin time; and (6) direction of ice movement.

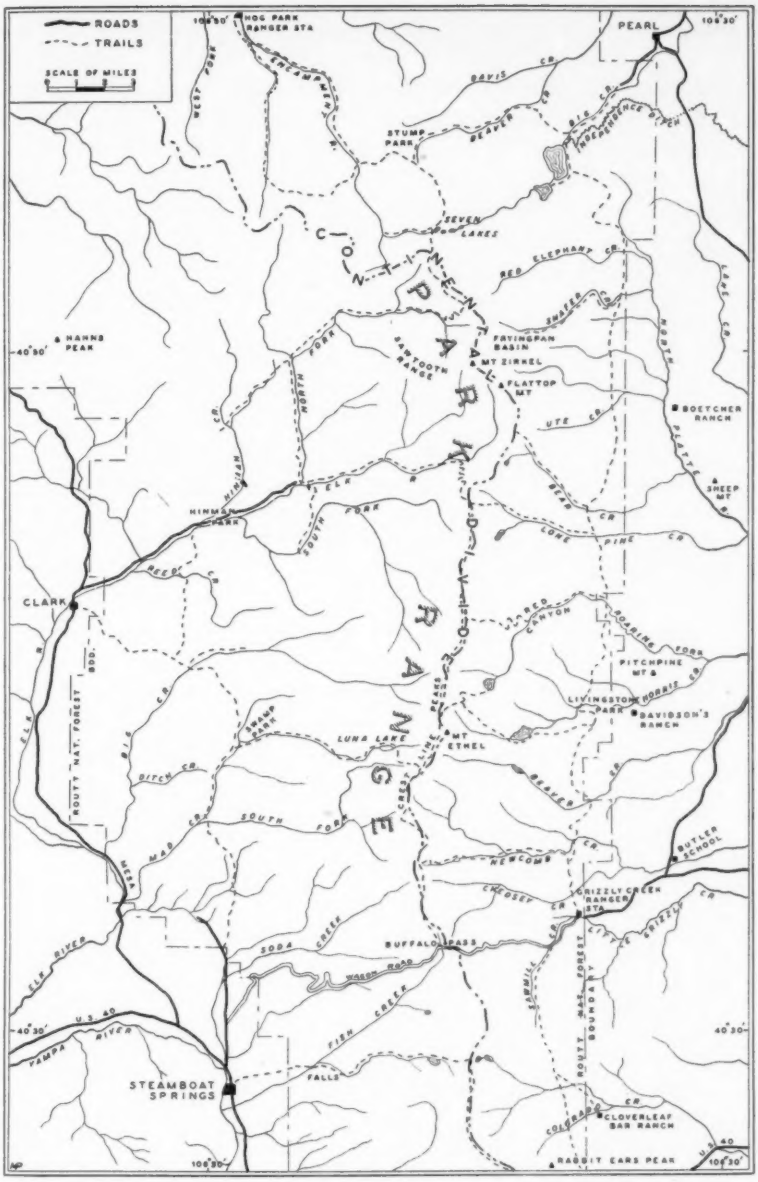


FIG. 10.—Index map of the Park Range. All places referred to in the description of Park Range glaciers are located on this map.

Within the catchment areas beautifully polished and striated surfaces abound and only scattered patches of moraine appear in this ice-scoured landscape. Not until the visitor reaches the western edge of Swamp Park does he find the typical frontal moraine with a hummocky topography and many undrained depressions. It is this accumulation of glacial drift that has blocked the drainage and produced Swamp Park.

Southwest of this locality a portion of Mad Creek Glacier forced its way across a low divide, into Ditch Creek. The moraine left by this small tongue of ice contains many large granite boulders 15–20 feet in diameter. It is possible that some of this material was left by earlier glaciers, and was only re-worked by Wisconsin ice. Farther to the southwest is the frontal moraine of the main body of Mad Creek Glacier. It is not well developed and fails to exhibit the usual hummocky topography associated with marginal deposits. Beyond this point, however, evidence of Wisconsin glaciation is entirely lacking.

On the map of glacial features (Fig. 9) one lateral and one medial moraine appear 3 miles northeast of the Mad Creek frontal. These are well-developed ridges and help to establish the upper limit of ice work in the valley. On the basis of these deposits, and other less perfectly formed moraines, the surface gradient of Mad Creek Glacier has been calculated to be approximately 280–300 feet per mile.

*Soda Creek Glacier.*—South of the crest-line peaks of the Park Range, the Continental Divide was ice-capped during Wisconsin time. Normal cirque development did not occur, and instead small ice-scoured depressions were formed in which lakes and swamps exist today. Beautifully polished and striated rock surfaces characterize the summit of the range where ice-cap conditions once prevailed (Fig. 12).

Supplied from the ice cap, several valley glaciers developed at the southern end of the range. One of these, Soda Creek Glacier, deserves special mention because of the tremendous lateral moraines which it produced. These moraines are clearly visible from the wagon road leading from Steamboat Springs to Buffalo Pass. The best-developed ridge follows the north side of the valley for more than 2 miles, terminating near the National Forest boundary $3\frac{1}{2}$

miles north of Steamboat Springs (Fig. 10). Outwash from Soda Creek Glacier covers most of the valley floor beyond the terminal moraines.

FIG. 11.—The east end of Hinman Park showing outwash in the foreground and the narrow valley of Elk River beyond. The upper limit of Wisconsin ice is indicated by the white line. The effect of ice scour is visible on the left side of the valley.

FIG. 12.—One of many exposures of glacial striae in the high country south of Mount Ethel. The hard granites and gneisses have retained glacial markings remarkably well.

*Fish Creek Glacier.*—Another valley glacier, fed by the ice cap, followed Fish Creek, which flows southwestward from Buffalo Pass. This mass of ice scoured the upper portion of the valley, leaving only scattered areas of heavy moraine and a thin veneer of drift between polished rock surfaces. At the lower end of the valley lateral moraines were left high on the walls (Fig. 13). In some places one, and in others two distinct laterals are present. Below these moraines polished surfaces of bed rock are common.

FIG. 13.—The ice-scoured valley of Fish Creek 3 miles east-northeast of Steamboat Springs. A lateral moraine stands out as a dark forested ridge on the north side of the valley. In the foreground are numerous polished granite surfaces.

The frontal moraine of Fish Creek Glacier is located on the lowland to the south of the range, and a portion of a curved morainic ridge is still intact. Postglacial dissection has altered many of the depositional features in this section and produced a small gorge above the falls of Fish Creek 3 miles east of Steamboat Springs. Outwash deposits are limited to a small area between the frontal moraine and the flood plain of the Yampa River. Most of the wash was presumably carried away by this stream.

*Rabbit Ears Glacier.*—Located at the southern end of the Buffalo Pass ice cap, Rabbit Ears Glacier marks the southern limit of Wisconsin ice in the Park Range. Although the frontal moraines of the glacier are south of the area shown on the map of glacial features,

the position of the ice is well pictured in Plate I. The most easily visited terminal deposits occur ½ mile west of Rabbit Ears Pass on United States Highway No. 40. Here the drift is composed of gray granite boulders and a variety of volcanic rocks, many of which are well striated. The hummocky topography of the moraine is visible from the road.

*Cloverleaf Glacier.*—East of the Continental Divide and south of Buffalo Pass is the valley of Colorado Creek, once occupied by Cloverleaf Glacier. Supplied with ice from the Continental Divide, this southeastern extension of the Buffalo Pass ice cap reached its maximum position at Cloverleaf Bar Ranch 2 miles northeast of Rabbit Ears Peak. Heavy moraines with hummocky topography and undrained depressions characterize the country west and northwest of the Ranch. Several swampy parklike areas occur within the morainic deposits.

North of the main body of Cloverleaf Glacier and south of Newcomb Glacier, ice pushed eastward from the Continental Divide and reached Sawmill Creek. Confronted by a high ridge just east of the Creek, the ice was diverted to the north and south. In consequence of this shift in direction, frontal moraines in this locality are closer to the range than elsewhere along the eastern front of the mountains (Fig. 9).

*Newcomb Glacier.*—With ice acquired from the large cirques south of Mount Ethel, and the Buffalo Pass ice cap, Newcomb Glacier pushed down the eastern slope of the Park Range to terminate on the lowlands of North Park at an elevation of 8,500 feet. The huge amphitheatral cirques and deeply grooved granite rocks give striking evidence of the erosive action of this ice mass. The abundance of débris gathered in the high country now forms the massive moraines of Newcomb valley. One of the best exposures of morainic topography is found south of the stream 2 miles north of Grizzly Creek Ranger Station. Other well-developed ridges follow the northern side of the valley and circle around to form the terminal moraine 2 miles west-northwest of Butler School. Although Newcomb Creek has cut through the frontal deposits and partially drained the region behind the moraine, a large parklike area still exists and serves as excellent summer pasture for cattle.

Immediately to the north of Newcomb Glacier two smaller tongues of ice issued from narrow mountain valleys. These glaciers received their ice from cirques south of Mount Ethel and then proceeded southeastward to the edge of the range where they terminated just north of Newcomb Creek. The moraines left by these valley glaciers merge with the larger drift deposits of Newcomb Glacier.

Beyond the frontal moraines, outwash material is common. Glacial waters followed Beaver, Newcomb, Chedsey, and Little Grizzly creeks, distributing large amounts of gravel which have since been veneered with recent stream alluvium. The extent of these wash deposits is indicated on the map of glacial features (Fig. 9).

*Livingstone Glacier.*—Slightly less than 6 miles in length and usually less than a mile in width, Livingstone Glacier is one of the smallest glaciers to originate on the slopes of Mount Ethel. With the possible exception of Red Canyon Glacier to the north, it can claim the most perfect system of moraines in the entire range. Having originated in a beautiful cirque on the eastern face of Mount Ethel, it followed the narrow valley of Norris Creek to the base of the mountains, where it piled up heavy moraines to heights of over 300 feet. Pink granite boulders 15–20 feet in diameter are common in the frontal deposits. On either side of the valley are high laterals which join with the terminal moraine to produce a large, horseshoe-shaped amphitheater.

During postglacial times Norris Creek has cut a narrow valley through the frontal moraine east of Livingstone Park. In spite of this lowering of the stream course, several small lakes located in the parklike valley still remain undrained. It is very likely that a larger lake once occupied this area.

North of Davidson's Ranch at an elevation a little below 8,500 feet the massive frontal moraine of Livingstone Glacier ends abruptly. Beyond this point outwash material occupies the low country south and east of Pitchpine Mountain. In the valley of Norris Creek well records indicate that 20 feet of gravel exist beneath the recent stream alluvium.

*Red Canyon Glacier.*—Generously supplied with ice from the Continental Divide north of Mount Ethel, Red Canyon Glacier scoured

and deepened the Valley of Roaring Fork, and extended almost 3 miles out onto the lowlands of North Park. In the high country polished and striated surfaces are abundant and only scattered patches of thin moraine give evidence of depositional work. In the canyon section ice reached a thickness of well over a thousand feet, as indicated by striae high on the walls. Where the ice issued from the mountains there is a beautiful U-shaped valley.

FIG. 14.—One of the large lateral moraines of Red Canyon Glacier viewed from a position on the terminal deposits 2 miles east of the entrance to Red Canyon.

Like the other glaciers which descended the eastern slope of the Park Range, Red Canyon Glacier produced heavy lateral and frontal moraines. From the road which leads to the mouth of the canyon these features are beautifully displayed. The lateral on the north forms an unbroken ridge approximately 2 miles in length. Viewed from the frontal moraine, this marginal deposit, which rises over 400 feet above the valley floor of Roaring Fork, appears as a giant railroad grade, terminating at the edge of the mountains (Fig. 14).

The frontal deposits consist of irregular hills separated by depressions, many of which contain small lakes or swamps (Figs. 15 and

16). The landscape is one of youth, and stands out in striking contrast to the mature glacial topography of the pre-Wisconsin moraines immediately to the east (Fig. 18). A narrow strip of outwash from Red Canyon Glacier follows the valley of Roaring Fork for several miles. On either side are low gravel terraces related to earlier glacial stages.

FIG. 15.—A portion of the Wisconsin terminal moraine of Red Canyon Glacier. Hummocky topography with many small lakes characterizes this perfect morainic landscape.

*Lone Pine and Bear Creek glaciers.*—Separated by a narrow ridge, Lone Pine and Bear Creek glaciers paralleled one another throughout their courses. Both glaciers originated in large picturesque cirques east of the Continental Divide and terminated a mile west of the North Platte River at 8,500 feet elevation. The steep cirque and valley walls and many highly polished rock surfaces give indication of the tremendous erosive work accomplished by these two glaciers. On the lowland east of the mountains heavy boulder moraines mark the maximum advance of the ice, while a series of recessional ridges record temporary halts in the melting back of the glaciers. From a

position on the North Platte River south of Sheep Mountain the frontal deposits appear as huge dams blocking the valleys of Lone Pine and Bear Creeks. To the east of these barriers outwash fills the valley of the North Platte. Near the stream the gravels have been partially covered by recent alluvium but close to the moraines the outwash material is at the surface.

FIG. 16.—A well-preserved terminal moraine north of Roaring Fork. These heavy drift deposits mark the maximum advance of Red Canyon Glacier during Wisconsin time.

*Mount Zirkel and Flattop Mountain glaciers.*—East of the Continental Divide in the vicinity of Mount Zirkel, small but well-formed cirques furnished ice to several short valley glaciers (Fig. 17). Three of these descended the eastern face of Flattop Mountain to terminate on the lowland a short distance away. Others followed Ute, Shafer, and Red Elephant Creeks.

Of the several glaciers, the one which originated in Fryingpan Basin, northeast of Mount Zirkel and then proceeded down the valley of Shafer Creek, was the largest and most effective. In addition to carving a beautiful cirque and steepening the valley walls, this mass of ice left a system of moraines east of the mountain front.

Outwash from the various glaciers followed the North Platte River for many miles to the south.

*Independence Glacier.*—The visitor who enters the Park Range from the northeast will very likely follow the Big Creek trail which leads from Pearl to the source of Independence Ditch. He will then bear to the north across a low divide and proceed up the valley of Beaver Creek. Just before making this shift, however, he will pass a huge morainic dam, the frontal deposit of Independence Glacier.

FIG. 17.—A cirque lake close to the Continental Divide north of Mount Zirkel. Ice from this area moved down the valley of Red Elephant Creek. Although no glaciers now exist in the Park Range, snow frequently remains in sheltered cirques until late in August.

If he should take a short detour to the top of this ridge he would see to the south of him two lakes, the product of glacial blocking. To the northeast would be the valley through which he had traveled and a narrow strip of outwash on either side of the stream. All around him would be the hummocky topography of heavy drift.

By nightfall this same traveler might pitch camp at Seven Lakes on the high divide between Big Creek and Encampment River. From here he could peer into the ice-polished and scoured valley once occupied by Independence Glacier. Like the ice of Red Canyon and Norris Creek, Independence Glacier has left a perfect record of its activities in the region.

*Beaver Glacier.*—At the northeastern edge of the Park Range two small glaciers left their imprint on the landscape. The southernmost has been called Beaver Glacier, while the other, located in the valley of Davis Creek, has not been named. From the trail which follows Beaver Creek some excellent moraines are visible, but the major area of frontal deposits is located downstream from where the trail enters the valley (Figs. 9 and 10). On account of heavy down timber, this section was traversed with great difficulty. However, over 2 square miles of hummocky topography with many lakes and swamps were observed.

*Encampment Glacier.*—The catchment areas of Encampment Glacier included the cirques on the north side of the Sawtooth Range (Fig. 8), the high divide near Seven Lakes, and the broad valley in which Stump Park is now located. Ice from these three sources found its way into the valley of Encampment River where it formed a glacier over 1,000 feet thick. Some of the ice pushed across a low divide to the west and followed West Fork, joining the main glacier southeast of Hog Park Ranger Station.

Scattered areas of heavy moraine in the valley of Encampment River indicate that the glacier made frequent halts during recessional stages. The trail which leads from Stump Park to Hog Park Ranger Station passes several of these deposits where hummocky topography is well exhibited.

Near the junction of Encampment River and West Fork a medial moraine separates the two valleys for a distance of more than a mile. This ridge, and the laterals to the northeast and southwest, mark the position of ice in the valley. To the northwest, where the frontal moraines should occur, there are only a few poorly defined hills of drift. In place of a morainic dam across the valley, there is a broad flat underlain by gravel outwash. Most of the débris carried by the Encampment Glacier was washed downstream instead of being deposited as heavy moraines like those associated with the glaciers on the east side of the range.

*Contrast between east- and west-slope glaciers.*—The glaciers which descended the west slope of the Park Range were longer than those on the east and terminated approximately 1,000 feet lower. Elk, Big Creek, and Mad Creek glaciers, for example, reached their

terminal positions between 7,000 and 7,500 feet, while all the east-slope glaciers terminated close to 8,500 feet. These differences are explained by the favorable exposure to moisture-bearing westerly winds.

Another contrast is seen in the character of the terminal moraines on the two sides of the range. On the east the deposits are large and well preserved, while on the west they are small and sometimes lacking. This is not the case with lateral moraines, however, for they are frequently well shown on the east and west. The poorly developed terminals are possibly due to more rapid melting at relatively low altitudes which produced swift streams that carried away large amounts of débris as outwash. This theory is supported by extensive wash deposits in the Elk River valley. On the east side, where terminal moraines are large, the amount of outwash is relatively small.

### PRE-WISCONSIN GLACIATION

The existence of several areas of old moraine, located beyond the limits of Wisconsin glaciation, indicate that ice occupied the valleys of the Park Range at least twice during the Pleistocene period. Some of the best evidence of the early ice appears in the form of massive moraines just east of the Wisconsin terminals of Newcomb, Red Canyon, and Lone Pine glaciers. Like the remnants of old drift in the Medicine Bow Range, these areas have a mature topography free of lakes and swamps. During the long period since this early glaciation, weathering has destroyed many of the rocks of the moraine. The larger boulders have been reduced in size while smaller ones have been broken into small, angular fragments. Today scattered weathered boulders are the chief features of the subdued landscape. Where the old and young moraines are side by side, the difference in appearance is very noticeable (Fig. 18).

From the map of glacial features (Fig. 9) it will be noted that each area of early drift is located just beyond the Wisconsin terminal moraines. This position signifies that pre-Wisconsin glaciers in the Park Range were only slightly larger than those which formed in late Pleistocene time.

On the west side of the range two remnants of the old moraine are worthy of note. The first of these is located south of Hinman

Park where Big Creek Glacier overflowed its own valley and pushed across a low divide into Reed Creek, a tributary of Elk River. The record of this movement is seen in mature morainic topography north of the Wisconsin lateral of Big Creek Glacier. The youthful appearance of some drift deposits within the Reed Creek area, however, suggests that possibly early Wisconsin ice also pushed into this section from Big Creek before the present lateral moraine was formed.

FIG. 18.—The gently rolling boulder-strewn surface in the foreground of the photograph is pre-Wisconsin drift. In the distance is a morainic ridge of younger age.

The second and most striking remnant of old drift appears on the "Mesa" north of the junction of Mad Creek and Elk River. Here, above and beyond all Wisconsin moraines is a collection of huge granitic boulders, some of them 40 feet across. They are foreign to the "Mesa" and therefore not the result of bed-rock weathering. Located outside the present valley, these boulders present a real problem in age determination. The possibility that they antedate the canyon cutting has been considered, but the absence of other deposits which could be related to a third and very early Pleistocene glaciation has caused the author to associate these boulders with the pre-Wisconsin valley glaciers, records of which have been found at several places within the region. The Park Range very likely ex-

perienced three periods of glaciation during the Pleistocene, but satisfactory evidence of the first advance was not discovered.

## CONCLUSION

As further studies of glacial features in the Rocky Mountain region are carried on, a more complex history than that yet discovered in the Medicine Bow and Park ranges should be anticipated. The author, in association with Wallace W. Atwood, Sr., has found, at several localities, evidence of Pleistocene ice which existed before the modern canyons were excavated. These ice fields were far more extensive than the valley glaciers described in this report. Alpine glaciers existed in the Rocky Mountains of Eocene time, and more evidence of that ice age should be found. Each heavy bouldery deposit discovered in the mountain area should be carefully examined for evidence of ice action. Each generation of Rocky Mountains probably contained Alpine glaciers.

# DEVELOPMENT AND STRUCTURAL CONTROL OF EROSION SURFACES

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ABSTRACT

Projected profiles and field observations indicate that there are three major and one minor upland surfaces in the Appalachian Plateau of western Pennsylvania and eastern Ohio. Arguments are advanced to prove that these represent erosion surfaces developed under four distinct baselevel controls. The effect of structure upon the development and configuration of these surfaces is considered.

## INTRODUCTION

The development, number, and correlation of the erosion surfaces in the Appalachian Highlands has been a subject of discussion since Davis recognized the Schooley peneplain.[1] Many of the earlier workers studied in the eastern portion of this area. As the structure is complex and a great number of formations of variable resistance appear on the surface, it is not surprising that a great number of surfaces were recognized and named.

Recently there has been a tendency to combine these surfaces, accounting for the differences in elevation by (1) the variable resistance of the rocks involved, (2) the distance from the baselevel of control, and (3) the fact that all the surfaces represent at the best only partial peneplains with pronounced highland and lowland areas which must be considered as integral parts of the same surface. Using these arguments, Ver Steeg[2] combined the various surfaces into three, namely, the Schooley (Kittatinny), Harrisburg, and Somerville.

The writer,[3] working in eastern and southern Ohio and adjacent areas of Pennsylvania and West Virginia, recognized four surfaces. Recently Ashley has presented the view "that the present surface,

[1] W. M. Davis, "The Rivers of Northern New Jersey, with Notes of the Classification of Rivers in General," *Natl. Geog. Mag.*, Vol. II (1890), pp. 81–110.

[2] K. Ver Steeg, "Erosion Surfaces of Appalachian Plateau," *Pan-Amer. Geol.*, Vol. LVIII (1932), pp. 31–44.

[3] W. S. Cole, "Identification of Erosion Surfaces in Eastern and Southern Ohio," *Jour. Geol.*, Vol. XLII (1934), pp. 285–94.

though reflecting a single old peneplane, has been lowered by not less than 100 feet for the hardest rocks and by several hundred feet for the softer rocks, for each million years since the beginning of uplift."[4] Thus, the lower surfaces of Ver Steeg, the writer, and others would result from this differential lowering, assisted and controlled by local baseleveling, stripping of flat-lying or nearly flat-lying resistant rocks, parallel lowering because of uniformity of rock and structure, stream captures or other factors of this type.

This idea injected into the controversy is refreshing and stimulating, and to date represents the extreme reduction in number of the Appalachian peneplains. The purpose of the present article is to test this view with regard to the data assembled by the writer. The Appalachian Plateaus offer the best area for such study because the structure is comparatively simple and detailed mapping has assembled a vast amount of information on rock resistance, drainage patterns, and other necessary data.

The index map (Fig. 1) shows the area over which detailed studies were made. Projected profiles and structure sections (Figs. 2–4) were constructed for certain areas not previously studied in detail. These are described below.

## DELAWARE-BRINKHAVEN SECTION

The summit elevations across the Delaware and western parts of the Marengo quadrangles (Fig. 2, top) are uniformly at about 1,000 feet. Near the center of the Marengo Quadrangle a steep rise is noted, which culminates on the Fredericktown Quadrangle. In this area the foreground summits have elevations of about 1,200 feet, while the background summits rise to 1,400 feet. Eastward there is a descent with the profile showing a more or less level surface. The area represented by this profile should be compared with the contour map drawn from hilltop elevations.[5]

White[6] has identified in the area represented by the western quadrangles of this profile a low plateau, which encircles an out-

[4] G. H. Ashley, "Studies in Appalachian Mountain Sculpture," *Bull. Geol. Soc. Amer.*, Vol. XLVI (1935), p. 1398.

[5] W. S. Cole, "Rock Resistance and Peneplain Expression," *Jour. Geol.*, Vol. XLIII (1935), p. 1052, Fig. 2.

[6] G. W. White, "The Powell and Broadway End Moraines," *Amer. Jour. Sci.*, Vol. XXX (1935), pp. 34–35, Fig. 2.

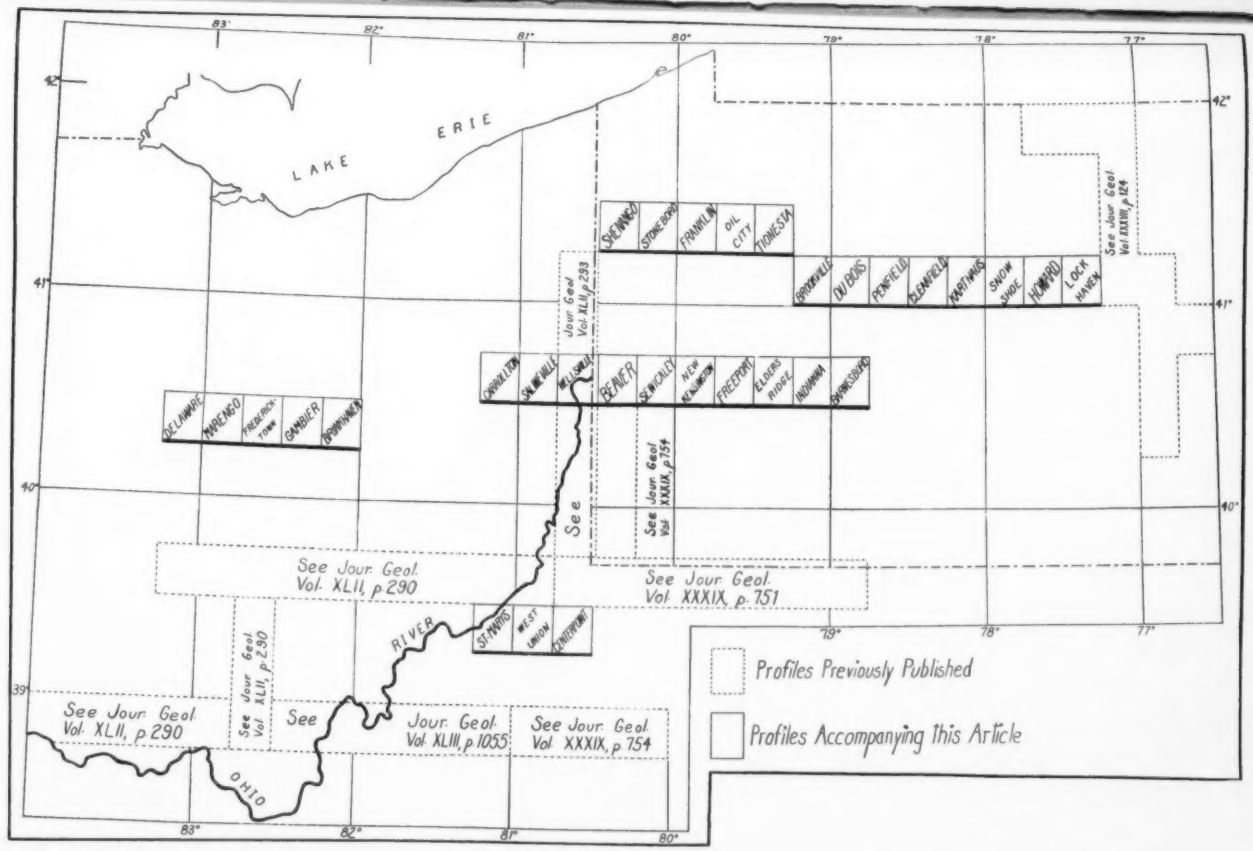


FIG. 1.—Index map, showing portions of Ohio, Pennsylvania, and West Virginia. The heavy line in each section indicates the base line of the projected profile and position of the structural section.

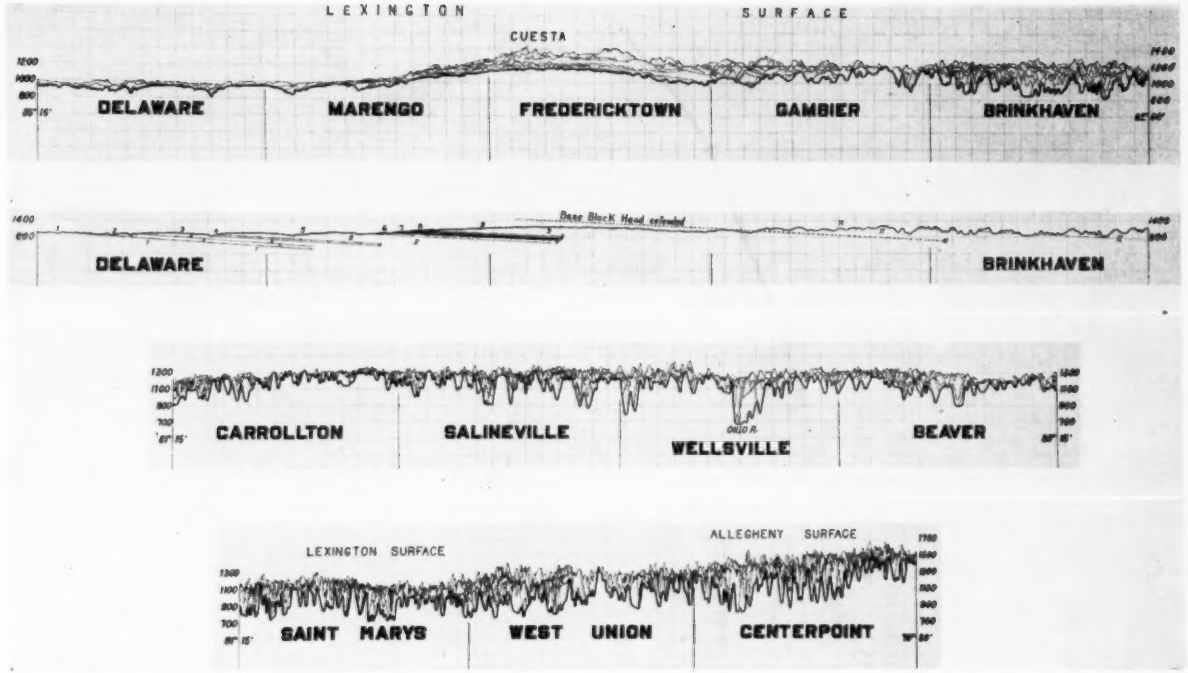


FIG. 2.—Top: projected profiles of the Delaware-Brinkhaven section. Upper center: structural section of the Delaware-Brinkhaven area showing relationship of resistant beds to topography. The formations represented are: (1) Monroe dolomite; (2) Columbus limestone; (3) Delaware limestone; (4) Olentangy shale; (5) Ohio shale; (6) Bedford shale; (7) Berea sandstone; (8) Sunbury shale; (9) Cuyahoga formation to base of Black Hand member; (10) base of Black Hand member; (11) Cuyahoga and Logan formations; (12) base of Pennsylvanian (Sharon conglomerate). Lower center: projected profiles of the Carrollton-Beaver section. Bottom: projected profiles of the St. Marys–Centerpoint section. Base lines and structural section along the southern boundary of the quadrangles. Vertical exaggeration of projected profiles is about $\times 26$; of the structural section, about $\times 8.7$.

lier of his "Allegheny" plateau. This outlier, which he names the Sparta outlier, is represented by the high elevations observed on the projected profile of the Fredericktown Quadrangle. The low plateau is a low, cuesta-like bench in front of the main Mississippian cuesta of central Ohio, and owes its position to the occurrence of the Berea sandstone. It is analogous to the situation described by the writer in the Morehead and Salt Lick quadrangles, Kentucky.[7] This interpretation had been given to this bench previously by Sharp.[8] The structure section (Fig. 2, upper center) clearly indicates the relationship between the topography and resistant beds. The low area east of the Sparta outlier marks the position of a preglacial valley.

Except for the occurrence of the minor Parker strath which in this area is masked generally by glacial deposits, only one erosion surface can be recognized, namely, the Lexington.[9] This bench as well as the main cuesta represent irregularities on this surface and should be considered as such.

### ST. MARYS–CENTERPOINT SECTION

The surface represented by this profile (Fig. 2, bottom) gradually rises eastward from about 1,200 feet to nearly 1,700 feet. The Lexington surface grades upward gradually to meet the Allegheny. The typical Allegheny surface starts at the western edge of the Centerpoint Quadrangle.

### CARROLLTON–BEAVER SECTION

The average elevations of the summits shown by this profile (Fig. 2, lower center) are slightly over 1,300 feet. The highest elevations are on the Wellsville Quadrangle where isolated areas rise above 1,400 feet. The major portion of the surface may be correlated with the Lexington, but the higher areas of the Wellsville Quadrangle represent outliers of the Allegheny surface.

### SEWICKLEY–BARNESBORO SECTION

From the western edge of the Sewickley Quadrangle (Fig. 4, top) with elevations slightly over 1,200 feet, there is a gradual rise east-

[7] Cole, "Rock Resistance and Peneplain Expression," *op. cit.*, p. 1060.

[8] H. S. Sharp, "The Geomorphic Development of Central Ohio," *Denison Univ. Bull. 27* (1932), pp. 20–23.

[9] Cole, "Rock Resistance and Peneplain Expression," *op. cit.*, p. 1059.

section, about ×8.7.

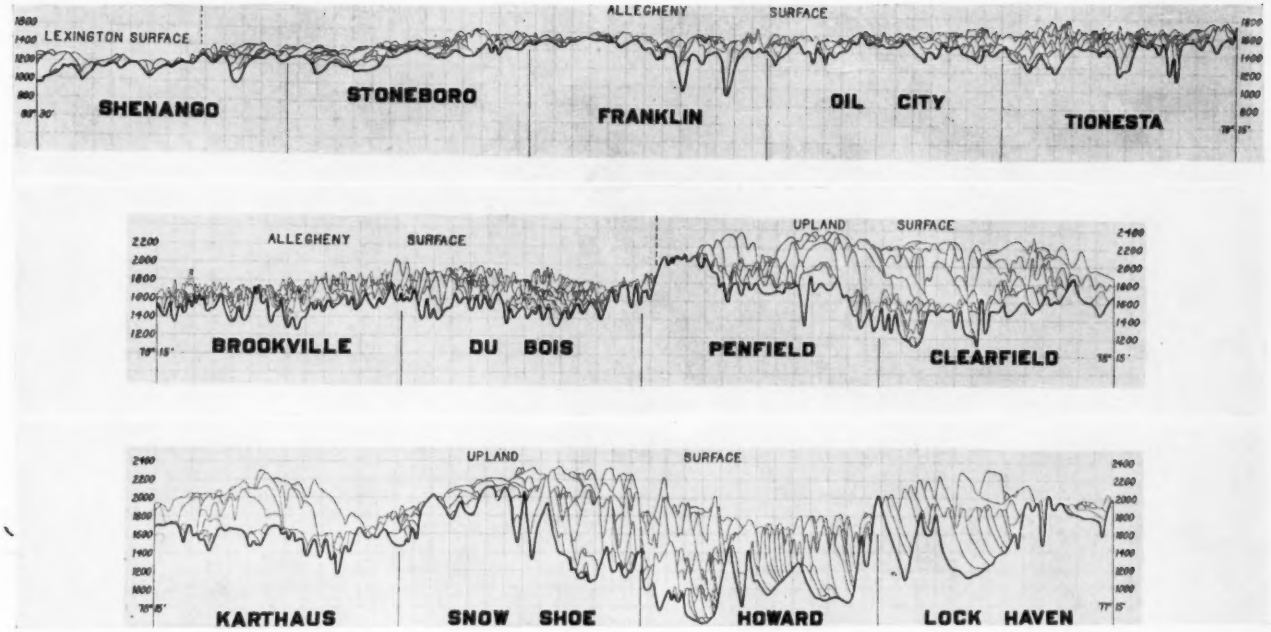


FIG. 3.—Top: projected profiles of the Shenango-Tionesta section. Center: projected profiles of the Brookville-Clearfield section. Bottom: projected profiles of the Karthausiz-Lock Haven section. Base lines along the southern boundary of the quadrangles.

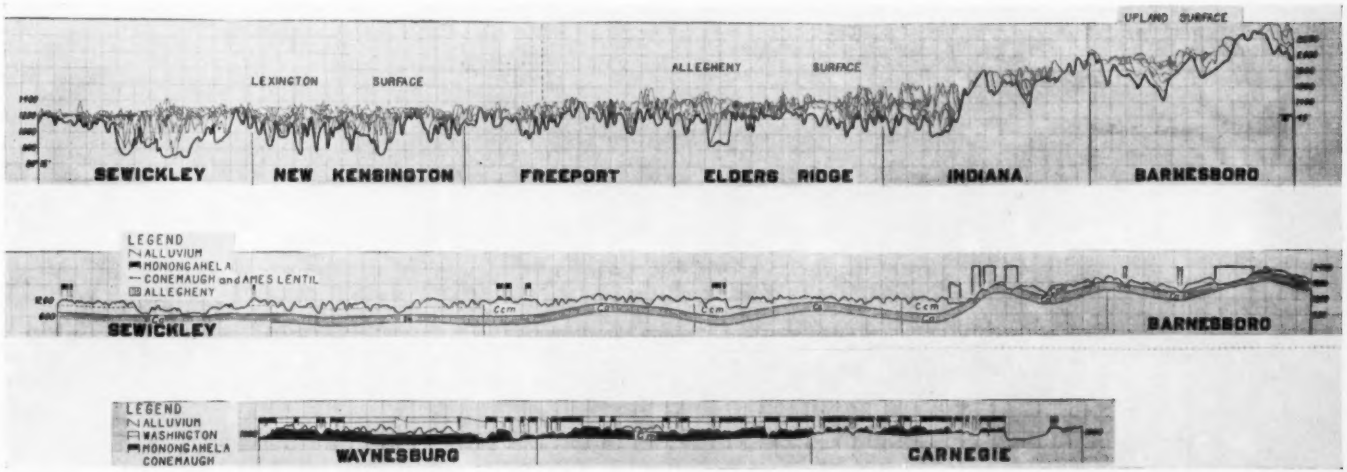


FIG. 4.—Top: projected profiles of the Sewickley-Barnesboro section. Center: structural section of the Sewickley-Barnesboro area. Base line of projected profile and structural section along the southern boundary of the quadrangle. Bottom: structural section of the Carnegie-Waynesburg area. Section drawn along the eastern boundary of the quadrangles.

ward to the central portion of the Freeport Quadrangle. In this area the elevations are about 1,400 feet. This surface continues eastward across the Elders Ridge Quadrangle, interrupted by certain isolated higher areas, some of which approach 1,600 feet.

In the extreme eastern portion of the Elders Ridge Quadrangle and extending through the western portion of the Indiana Quadrangle, certain summit elevations of the northern portion are slightly above 1,600 feet, although the summits in the southern portion are at about 1,400 feet. This is followed by a slight break with the elevations increasing to over 1,800 feet. Structural control plays an important part in the formation of this break. This slightly higher area must be correlated with the lower portion shown on the profile west of this area. Eastward there is a relatively rapid increase in elevation, so that at the eastern edge of the Barnesboro Quadrangle certain areas have elevations of about 2,400 feet.

Ashley has drawn contour maps from the hilltop elevations which present in plan view the area here profiled. Reference should be made to his general contour map[10] of Pennsylvania and adjoining areas and to his earlier maps showing the Schooley and higher (?) peneplains[11] and the Allegheny surface.[12]

The summit area between 1,200+ feet and 1,400 feet represents a re-entrant eastward of the Lexington surface[13] which is more perfectly developed in eastern Ohio and adjacent areas. The surface between about 1,400 feet and 1,800 feet represents the Allegheny, while the higher surface on the eastern portion of the profile represents the Upland erosion surface.

The sequence of rocks exposed at the surface in the area here profiled ranges from Mauch Chunk (Upper Mississippian) to Monongahela. The greater portion of the area, however, has rocks of Conemaugh age at the surface. Figure 4, center, represents a structure section drawn along the same line as the one selected for the base

[10] Ashley, *op. cit.*, p. 1400, Fig. 1.

[11] G. H. Ashley, "The Scenery of Pennsylvania," *Pa. Top. and Geol. Surv., Bull. G6* (1933), p. 19, Fig. 26.

[12] *Ibid.*, p. 24, Fig. 30.

[13] W. S. Cole, "Identification of Erosion Surfaces in Eastern and Southern Ohio," *Jour. Geol.*, Vol. XLII (1934), pp. 285–94; and "Rock Resistance and Peneplain Expression," *ibid.*, Vol. XLIII (1935), pp. 1049–62.

line of the projected profile. The structure of the entire area is a series of gentle folds, but those of the western portion are so low that the rocks are practically horizontal. The folds increase in intensity eastward. Casual inspection at once indicates an apparent relationship between structure and topography along this profile.

### SHENANGO-TIONESTA SECTION

From the western edge to a point just east of the central portion of the Shenango Quadrangle (Fig. 3, top), the elevations vary from about 1,200 feet to a maximum of 1,300 feet. Eastward there is a progressive increase in elevation until 1,600 feet is reached at the eastern edge of the Stoneboro Quadrangle. This elevation is maintained eastward across the Franklin Quadrangle and the greater portion of the Oil City Quadrangle. At this place there begins a gradual but progressive increase in elevation, interrupted by a few isolated higher areas. The elevations approach 1,800 feet at the eastern edge of the Tionesta Quadrangle.

The Lexington surface is present in the western portion of the Shenango Quadrangle. The remaining quadrangles of this section exhibit the Allegheny surface, which slopes gently westward for the most part.

### BROOKVILLE-CLEARFIELD SECTION

From the western edge of the Brookville Quadrangle (Fig. 3, center), the Allegheny surface extends eastward to the western edge of the Penfield Quadrangle. The profile indicates that the surface increases in elevation eastward from slightly below 1,800 feet to nearly 2,000 feet in the western portion of the Du Bois Quadrangle. Eastward from this there is a slight decrease in elevation.

At the western edge of the Penfield Quadrangle an abrupt increase in elevation is noted. These higher elevations, maintained by certain summits, particularly in the background, extend to over 2,200 feet.

Beside the higher surface, there is a lower one, at about 1,600 feet, maintained by summits of the base line and adjacent to this line. This is especially well developed at the western edge of the Clearfield Quadrangle.

The surface maintained by the higher elevations through this por-

tion of the profile is the Upland, while the lower surface showing in the foreground is the Allegheny and correlates with the more extensive surface in the western portion of the profile.

KARTHAUS-LOCK HAVEN SECTION

Two surfaces are present on this profile (Fig. 3, bottom), an irregular, well-dissected one ranging in elevation from about 2,000 feet to over 2,300 feet, and one with elevations from about 1,400 feet to about 1,600 feet. The upper surface represents the Upland and the lower the eastward continuation of the Allegheny.

It should be noted that the boundary between the Appalachian Plateau province and the Valley and Ridge province is located near the western edge of the Howard Quadrangle. As this boundary trends northeast across both the Howard and Lock Haven quadrangles, the projected profiles of these quadrangles contain hilltops of both provinces. In each case, however, the base line and hills adjacent to it are those of the Valley and Ridge province.

The Lock Haven Quadrangle is just west of the Williamsport Quadrangle. Projected profiles of the latter quadrangle and quadrangles north and south of it have been published by Fridley[14] and Ver Steeg.[15] These authors present different interpretations of their profiles of this area. Fridley identifies four erosion surfaces on his combined Williamsport-Milton projected profile, namely, Kittatinny, Schooley, Honeybrook, and Harrisburg. In the same area Ver Steeg recognizes only two surfaces, the Schooley (Kittatinny) and Harrisburg.

The Upland surface[16] of the present profiles is the equivalent of the Schooley (Kittatinny) surface of Ver Steeg and the Kittatinny of Fridley. Below this the Allegheny surface may be observed. This apparently is to be correlated with the Schooley surface of Fridley and in part with the Harrisburg of Ver Steeg, inasmuch as Ver Steeg correlates the surface at about 1,600 to 1,700 feet on the Trout Run

[14] H. M. Fridley, "Identification of Erosion Surfaces in South Central New York," *Jour. Geol.*, Vol. XXXVII (1929), p. 124, Fig. 5.

[15] K. Ver Steeg, "Some Features of Appalachian Peneplains," *Pan-Amer. Geol.*, Vol. LIV (1930), Fig. 2.

[16] Cole, "Rock Resistance and Peneplain Expression," *op. cit.*, p. 1054.

Quadrangle[17] with the Harrisburg, while Fridley identifies it as Schooley. Fridley and Ver Steeg both recognize the Harrisburg as the youngest surface in this area. The present profiles do not show this to any extent.

### ANALYSIS OF THE PROJECTED PROFILES

Analysis of the projected profiles here presented, together with those previously published,[18] indicates that there are three major definite and recognizable surfaces and one minor surface in the area under discussion. These surfaces have been named the Upland, the Allegheny, the Lexington, and the Parker strath. The only one not shown with clarity is the Parker strath, because of mechanical difficulties inherent to projected profiles. Inspection of any number of quadrangles[19] within the area will, however, reveal the presence of this strath.

Are these surfaces the result of four uplifts with periods of partial peneplanation intervening, or are they the product of the various processes and factors listed by Ashley?[20] In any case, whether the concept of repeated peneplanation is accepted or not, it is a generally recognized fact that areas of resistant rock and those portions of the terrain which are farthest from the major drainage lines will be the last to be reduced. The structure section and projected profile of the Sewickley-Barnesboro section (Fig. 4, center) reveal an apparent relationship between structure and topography. The structure sections of the Accident-Grantsville and the Masontown-Uniontown folios exhibit the same relationships. The problem, therefore, resolves itself into a proof of the existence of more than one surface unless these lower surfaces are to be explained by reduction from one peneplain by various factors and controls rather than a series of uplifts with partial peneplanation.

[17] Ver Steeg, "Some Features of Appalachian Peneplanes," *op. cit.*, p. 17; Fridley, *op. cit.*, p. 117, Fig. 2.

[18] H. M. Fridley and J. P. Nolting, Jr., "Peneplains of the Appalachian Plateau," *Jour. Geol.*, Vol. XXXIX (1931), pp. 751, 754; W. S. Cole, "Identification of Erosion Surfaces in Eastern and Southern Ohio, *ibid.*, Vol. XLII (1934), pp. 290, 293; and "Rock Resistance and Peneplain Expression," *ibid.*, Vol. XLIII (1935), p. 1055.

[19] Waverly and Sciotoville, Ohio; Milton and Saint Albans, West Virginia; Beaver and Carnegie, Pennsylvania.

[20] "Studies in Appalachian Mountain Sculpture," *op. cit.*, p. 1398.

### RESISTANCE OF THE ROCKS

In the Paleozoic series which is involved in the area, the Chemung, Catskill, Pocono, and Pottsville may at once be designated as the most resistant. These formations form the ridges and cap the higher uplands. The formations from the Allegheny to the Washington are practically the same so far as resistance is concerned, but many local variations occur. In the main, these are much less resistant than the Pottsville and others placed in the most resistant group. The Middle and Lower Devonian and Silurian of most of the area represent the rocks of least resistance.

The Upland surface has its major development on rocks of the more resistant type. It may be noted, also, that this area in general forms the divide region and is therefore less susceptible to reduction. The cuesta character of the prominent upland which extends across Ohio has been demonstrated to be due to the occurrence of gently east- and southeast-dipping resistant beds of Mississippian age with weak Middle and Lower Paleozoic formations outcropping to the west of these, and less resistant Upper Paleozoic beds on the east.[21]

From this it might be concluded that the entire area of the Appalachian Plateaus under consideration in this article owes its configuration to the position and resistance of the various formations involved. With one bold sweep, the various peneplains could thus be eliminated! That the structure of a given area plays an important part in the development of our landscapes cannot be gainsaid. But that it is the major control is a debatable point.

### PROOF OF SEPARATE EROSION SURFACES

Inasmuch as the younger surfaces are the best preserved, the detailed analysis should begin with these. It is well recognized that the present drainage of this area is largely the result of the glacial interruption of the preglacial system, with consequent diversions and reversals. The birth of the present Ohio River was contemporaneous with the first glaciation, following ponding, diversion, and integration of a number of former northwest-flowing streams.

The present valley floors are cut from a few feet to several hundred below the floors of the former valleys. This condition may be

[21] Cole, "Rock Resistance and Peneplain Expression," *op. cit.*, pp. 1056–59.

the result of uplift, but Mallott has adequately accounted for the major portion of the incision by static rejuvenation.[22]

The floors of the preglacial valleys, where preserved, indicate that these valleys had reached maturity before the interruption described above occurred. To the incipient surface represented by these valley floors, the name Parker strath is applied. The Parker strath occurs from a few feet to several hundred feet below the general upland level. That the streams which carved the Parker strath were rejuvenated streams, superimposed from terrain which was well advanced in the topographic cycle, can be definitely proved by the following characteristics, which are shown by nearly all of these valleys: (1) the long meander swings; (2) the lack of structural adjustment; (3) the apparent uniformity with which each independent system has cut below the upland surface; (4) the restored surface of Lexington time sloping uniformly in most places in the direction of the Parker strath valleys; and (5) the fact that each system is apparently controlled by the same baselevel.

Uplift of the Lexington erosion surface with the rejuvenation of the stream systems is the most logical explanation for these characteristics of the Parker strath valleys. Capture or the introduction of local baselevels will not account for the observed phenomena.

The contour maps of hilltop elevations constructed by Ashley[23] and the writer[24] should be compared with maps which show the position and direction of Parker strath streams.[25] The intimate relationship between the slopes shown by the contour maps and position of the streams is at once apparent. If an areal geologic map is introduced, it will be noted that the streams are not controlled by structure.

Projected profiles, previously published, indicate a distinct break between the Lexington and Allegheny surfaces. This may be seen

[22] C. Mallott, "Static Rejuvenation," *Science*, N.S., Vol. LII (1920), pp. 182–83; "The Physiography of Indiana," *Handbook of Indiana Geology*, Part II (1922), p. 169.

[23] "Studies in Appalachian Mountain Sculpture," *op. cit.*, p. 1400.

[24] "Rock Resistance and Peneplain Expression," *op. cit.*, p. 1052.

[25] S. Happ, "Drainage History of Southeastern Ohio and Adjacent West Virginia," *Jour. Geol.*, Vol. XLII (1934), p. 268, Fig. 3; Ashley, "The Scenery of Pennsylvania," *op. cit.*, p. 35, Fig. 37.

in the projected profiles of the New Martinsville Quadrangle[26] and the Arnoldsburg Quadrangle.[27] This break is not so distinct on the St. Marys–Centerpoint section (Fig. 2, bottom). The lower surface apparently grades into the upper. Local variations in resistance account for the differences exhibited.

While valley patterns do not assist in the interpretation of the Allegheny surface so much as in the younger ones, certain inferences may be drawn. Examination of Ashley's map[28] shows a marked low area extending along the present course of the Monongahela River from West Virginia to Pittsburgh, Pennsylvania, and thence northwestward into Ohio. This is the drainage area of the preglacial Beaver. As this preglacial stream inherited its course from the Lexington surface, the lowland is a re-entrant of the Lexington surface from Ohio into the Allegheny surface, which is maintained by uplands on either side of the re-entrant.

The structure section (Fig. 4, bottom) of the Carnegie–Waynesburg area should be compared with the projected profile published by Fridley and Nolting.[29] This shows the truncation of beds of the Washington, Monongahela, and Conemaugh formations. The slope of the general surface is northward toward the re-entrant of the Lexington surface in this area.

North of the re-entrant, the Conemaugh and Allegheny appear at the surface. The elevations in this area match those to the south. The slopes are also toward the old Beaver River.

These facts offer additional proof that the dominant control is the major drainage lines and that progressive reduction was taking place from the major streams toward the upper surface from which the lower was being formed.

### SEQUENCE OF EVENTS UNDER STREAM CONTROL

If rejuvenation of a given region occurs, first the main streams and then the tributaries will be stimulated. In a relatively short

[26] Cole, "Identification of Erosion Surfaces in Eastern and Southern Ohio," *op. cit.*, p. 290.

[27] Cole, "Rock Resistance and Peneplain Expression," *op. cit.*, p. 1055.

[28] "The Scenery of Pennsylvania," *op. cit.*, p. 24.

[29] *Op. cit.*, p. 754, Fig. 4.

time trenching will be accomplished throughout the region thus affected.[30] But general lowering of the area is not so much a function of the streams as it is of weathering and sheet wash. In the development of a region during a normal erosional cycle, the streams after their initial trenching serve mainly as transporting agents for the débris which is moved down the slopes by sheet wash. By lateral planation, the valleys will be widened, but the general reduction toward the peneplain is accomplished mainly by divide lowering. Therefore, except for their importance in trenching during youth, the streams assist the mature and old-age stages of land-form development mainly in their ability to transport the débris which they acquire, in most cases passively.

The present Ohio and its tributaries admirably illustrate the point. The Ohio was formed by the ponding and divergence of a number of northwest-flowing streams, which not only broke through the divides but cut the floor of the new valley thus formed from one to several hundred feet below the former level of drainage, developing throughout the area a new graded profile.

In this connection, it is of interest to note that the drainage which preceded the present had a graded course in all areas of detailed study. This drainage pattern is preserved as benches and abandoned valleys throughout the Appalachian Plateaus, one of the major abandoned valleys of this system being the Teays of southeastern Ohio.

The Teays stream, in order to have the course indicated by the abandoned valley, must have wandered widely across the vale east of the Mississippian cuesta. After rejuvenation, with the incision of the valley, the entire system became graded. This was followed by lowering of the divides, which was accomplished more rapidly in belts of weaker rock. Thus the Teays valley (Parker strath) of south-central Ohio does not correspond to the axial position of the lowland, as the valley is found in a marginal position. This may be

[30] This assumption is based on present conditions within the Appalachian Plateaus. It is not to be assumed that trenching will proceed as easily in areas such as those described by Wright ("The Older Appalachians of the South," *Denison Univ. Bull. 26* [1931], pp. 164–66) where local baselevels may be established for long enough periods to produce high-level partial peneplains.

checked by tracing the restored Teays system[31] on the map[32] constructed by the writer.

If the Parker strath system had not been interrupted, a wide, low surface would have been developed adjacent to the streams in belts of weak rock over the entire area. This surface would be definitely controlled by the gradient of the streams in the immediate vicinity.

### STRUCTURE AND THE SURFACES

The Upland surface is preserved because of two basic structural factors: (1) the resistance of the rocks; and (2) the distance from the baselevels which controlled the development of the various surfaces. If there had not been resistant rocks near the headwaters of the various drainage systems, this surface would probably have been completely destroyed. As it is found today, it is a fragmentary surface maintained only by these special structural factors.

Ashley has published a portion of the Penfield, Pennsylvania, Quadrangle to show the contrast between the high, flat surface maintained by Pottsville sandstone, at about 2,300 feet, in two anticlines and a lower surface between 1,600 and 1,800 feet developed on the Allegheny and Conemaugh formations. He interprets these to be parts of a single surface derived from an old peneplain, separated at the present time as the result of differential erosion.

If differential erosion is the only factor in the formation of these surfaces, it is difficult to account for the accordance of this lower surface with the extensive Allegheny surface just west of this quadrangle. It is more difficult to account for the uniformity of the lower surfaces (Allegheny and Lexington), if a definite control is not established for each. Differential erosion from one surface would produce more irregularity. It may be also argued that the slopes toward the major streams would not be so perfectly in accord if there had not been periods of partial peneplanation following each rejuvenation.

### EVIDENCE OF WARPING

The writer is in accord with Ver Steeg[33] that the surfaces show little evidence of uneqal warping. The profile slopes are uniform, as

[31] S. Happ, *loc. cit.*

[32] Cole, "Rock Resistance and Peneplain Expression," *op. cit.*, p. 1052.

[33] "Some Features of Appalachian Peneplains," *op. cit.*, p. 28.

can be observed from those presented in this paper. Any irregularity can be explained satisfactorily by differential sculpture or distance from the major drainage lines.

## AGE OF THE SURFACES

There is little evidence in the area studied to warrant definite statements concerning the ages of the various surfaces. However, the Parker strath is definitely preglacial, inasmuch as its development was checked by the advance of the Pleistocene ice sheet.

Kay has computed the minimum age for the beginning of the glacial period in Iowa as about 700,000 years.[34] This is a minimum figure as he states that "it would seem safe to state that the Pleistocene involved probably a million years, possibly twice this length of time." Thus, it is evident that the hypothesis presented by Ashley[35] that the Schooley (Upland of this article) surface may be of late Pliocene age is not in accord with the facts or history of this area. Moreover, the two recent and contrasted theories[36] advanced to explain the origin of the drainage of the eastern Appalachians require a greater age for the Upland surface than that allowed by Ashley.

## CONCLUSIONS

There are three major and one minor erosion surfaces in the Appalachian Plateaus in the area studied. The preservation of the oldest surface depends upon the resistance of the rock which underlies it, and the fact that this area has been and continues to be the divide region.

Study of the younger surfaces indicates that the major uplands on these surfaces are controlled by rock resistance, but the general slopes are directly related to the gradients of the drainage systems.

As the Parker strath can be proved definitely to be preglacial in its development, the older surfaces must be more ancient than Pliocene.

[34] G. F. Kay, "Classification and Duration of the Pleistocene Period," *Geol. Soc. Amer. Bull. 42* (1931), p. 466.

[35] "Studies in Appalachian Mountain Sculpture," *op. cit.*, p. 1436.

[36] D. Johnson, *Stream Sculpture on the Atlantic Slope* (New York: Columbia University Press, 1931); H. A. Meyerhoff and E. W. Olmstead, "The Origins of Appalachian Drainage," *Amer. Jour. Sci.*, Vol. XXXII (1936), pp. 21–42.

# A VIEW OF MAGMATIC DIFFERENTIATION

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ABSTRACT

This paper was delivered at the Tercentenary Conference of Arts and Sciences at Harvard University, September, 1936. It is a brief presentation of some of the major difficulties that are encountered in attempting to explain all the phenomena of differentiation of igneous magmas by a single process. Special attention is given to the theory of differentiation by crystal fractionation.

The attitude that a person takes toward a problem depends upon his background of experience. Dr. Bowen's approach to the problem of magmatic differentiation has been chiefly through laboratory experimentation with artificial melts, and he has brought to this a skill and knowledge of chemical principles that everyone must admire. The test of conclusions, however, lies with their applicability to natural phenomena, and from this standpoint I have been impressed with certain difficulties regarding the theory of crystal fractionation that appear to me to be serious.

The reliance placed upon laboratory experiments implies a belief that the "dry" melts of the laboratory have such a close resemblance to magmas that the results are directly applicable. At this early stage our views begin to diverge. Undoubtedly certain principles are of such fundamental character that they are applicable to all melts, but these principles are so general that they do not carry us far, and observation of magmatic phenomena has impressed me with the view that some of their most important characteristics are due to factors that have not been taken into account in laboratory experimentation.

Of course, everyone recognizes these influences in some degree. It is common knowledge, for instance, that pressure and the presence of volatiles are important in determining whether the ferromagnesian silicates will crystallize as one of the various pyroxenes or as hornblendes or micas. Now, among these several possibilities, which one prevails in the region of the lithosphere where the most important part of crystal fractionation is supposed to occur? Obviously, the

composition of the residual liquid will vary greatly according to the ferromagnesian mineral that separates.

It may be said that we cannot restrict the choice to only one of these minerals, we should recognize that any one of them may separate according to conditions; but if this allowance is made, the door is thrown open to so many possibilities as to the course differentiation may take that a wide variety of results may seem to be in accord with the theory. Perhaps it is partly because of this lack of limitations that so many petrologists have thought that all variations in igneous rocks might be ascribed to this process. In order to arrive at a reasonable appraisal of the validity of the theory, we shall have to inquire into some of its fundamental features.

In many masses of diabase we find the interstices among the larger crystals to be occupied by a micropegmatite of quartz and alkali feldspar. To this intergrowth great importance has been attached. It has been claimed to demonstrate the formation of a final residual liquid of approximately granitic composition, as the theory requires. On the other hand, there is reason to doubt whether these micropegmatites really represent magmatic rest-liquid. Without dwelling upon this, I wish to call attention to another feature found in many diabases, of which the origin is less controversial but of which the significance has received little attention. I refer to the dikelike, or irregular, masses of coarse basic pegmatite, such as those found in the Goose Creek diabase in Virginia that Shannon[1] described (Fig. 1).

These diabase pegmatites are naturally to be attributed to the segregation of residual magma after most of the mass has crystallized. In the Goose Creek quarry they form probably less than 5 per cent of the whole. In the Palisade diabase similar pegmatitic segregations occur, even more basic. They are a fairly common feature of diabases. Shannon regarded the Goose Creek pegmatite as the final residue of the magma, in which water and other volatiles had become concentrated, and, curiously, he seems to have believed that the occurrence supported Bowen's theory of crystal fractionation.

[1] E. V. Shannon, "The Mineralogy and Petrology of Intrusive Triassic Diabase at Goose Creek, Loudoun County, Virginia," *Proc. U.S. Natl. Mus.*, Vol. LXVI, Art. 2 (1924).

It may indeed be due largely to crystal fractionation, but the results are in opposition to those assumed by the theory. These final residues should have had the composition of a granite or granodiorite, instead of one obviously very basic.

We have, then, in this one body of diabase the quartzofeldspathic micropegmatite and the very basic macropegmatite. Unless we grant liquid immiscibility, both cannot represent residual magmas,

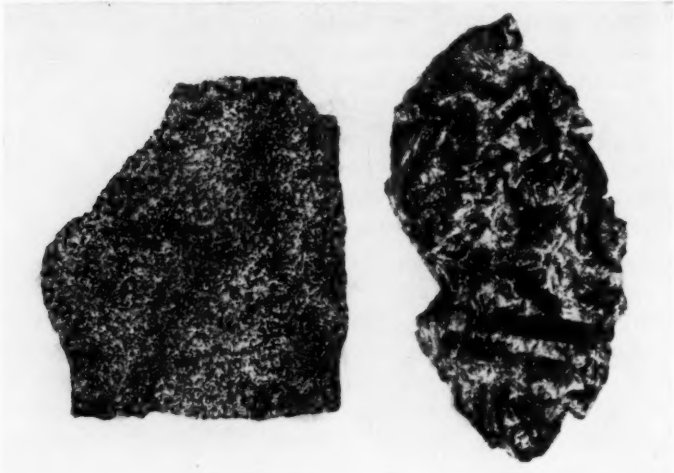


FIG. 1.—Diabase specimens from Goose Creek, Va. Left-hand specimen is the normal fine-grained diabase; right-hand specimen is the pegmatitic facies. The latter consists principally of bladelike pyroxene and feldspar. ×0.57.

for that would mean that the magma was differentiating simultaneously in two very different directions. On this fundamental matter the necessity of more adequate consideration is apparent.

An occurrence that has had much weight, especially among British petrologists, is that of the Glen More ring dike, described in the Mull Memoir.[2] At a number of places along the course of this dike exposures show gabbro below and granophyre above, though in certain exposures very anomalous associations are found. The authors

[2] H. H. Thomas and E. B. Bailey, "Tertiary and Post-Tertiary Geology of Mull," *Mem. Geol. Surv. Scotland* (1924), p. 330.

believe that after most of the gabbroic minerals had crystallized out of the original basic magma, there was a pause in crystallization over a range of temperature, during which a residue of interstitial granitic liquid filtered upward because of lighter specific gravity and formed the granophyre.

This explanation has been sponsored by some of the ablest British geologists, and it has been widely accepted by their colleagues. It has even been regarded as a demonstrated fact and as a standard by which to judge the validity of views proposed to explain relations in other localities; yet, as Arthur Holmes has clearly shown in a recent article,[3] the process violates fundamental principles of chemistry and dynamics. For example, the pause of crystallization over a range of temperature is something unknown in crystallizing solutions, and we have every reason to believe that it is impossible. Again, the hypothetical granitic residue would not have amounted to more than 5 or 10 per cent at most, and a much greater deforming force than gravity would have been required to squeeze it out of the interstices of the nearly solid gabbro. Furthermore, the explanation as a whole is not in accord with the theory of crystal fractionation that it purports to follow. The theory requires that in the passage from a gabbroic magma to a granitic, the diminishing liquid portion should pass successively through all intermediate compositions. Thomas and Bailey do not mention this requirement, nor do they make any allowance for these intermediate liquids.

The requirement of the theory that intermediate liquids of much greater bulk than granite must be formed cannot be escaped. Nevertheless, the curious phenomenon of repeated alternate outpourings of basalt and rhyolite with little or no intermediate magma is of not uncommon occurrence. It has attracted considerable attention, but the difficulty of reconciling it with crystal fractionation does not seem to have been generally realized.

The two magmas are evidently closely related in some manner and are affected by the same eruptive forces, as their successive outpourings follow each other during a short igneous cycle and sometimes are almost simultaneous. The phenomena seem to require the coexistence of basaltic and rhyolitic liquids. Has the rhyolite been derived from

[3] "The Idea of Contrasted Differentiation," *Geol. Mag.*, Vol. LXXIII (1936), p. 228.

the basalt? How has this been accomplished without the production of intermediate magmas? What is the relative position of the two magmas in the crust, and by what disposition of outlets does each reach the surface without contamination with the other?

In Yellowstone Park the basalts that alternated several times with rhyolites during the latest igneous epoch are described by Iddings as being almost free from phenocrysts.[4] This is significant. Since the basalts had not crystallized to any important degree before eruption, the rhyolites could hardly have been derived from them by crystal fractionation.

The view that we know enough about magmas to explain in principle all their phenomena and that, if crystal fractionation is not accepted, some alternative explanation of differentiation should be ready at hand has been implied in questions I have been asked at times. They take the form, "How *would* you explain differentiation?"

One may suggest several possible processes, and maintain with confidence that removal of mineral constituents in a vapor phase is of importance. Nevertheless, I do not feel that any process or combination of processes that has been formulated is competent to explain all phenomena. Possibly we shall not have a complete explanation until we know the ultimate sources of magmas and the causes of igneous activity. In response to the query that implies that crystal fractionation should be accepted unless a better explanation can be offered, I will propound a query in return: How is the differentiation of the Bushveld lopolith to be explained?

The general features of this enormous body of wonderfully differentiated rock are too well known to need description. One who studied it most carefully was the late Percy Wagner. In an early paper he attributed the differentiation to crystal fractionation. It cannot be supposed, therefore, that he was antagonistic to the idea. In 1929, shortly before his death, his book, *Platinum Deposits and Mines of South Africa*, was published. In it he describes the phenomena, discusses crystal fractionation and various other processes that might be suggested, and shows the apparent impossibility of all of

[4] J. P. Iddings, *U.S. Geol. Surv. Mono. 32*, Part II (1899), p. 433.

them. He admits without reservation that he does not know how the relations have been brought about. I quote:

> Here, however, we are entirely in the realm of speculation, and we must conclude this section by admitting reluctantly that no adequate explanation can at present be given of the remarkable features presented by the Differentiated Zone of the Bushveld Complex.[5]

F. E. Wright, who was a member of the Shaler Memorial Expedition to South Africa, made a careful study of the Bushveld complex. He is equally certain that it is out of the question to explain the relations by crystal fractionation.

I suggest as something to ponder over the fact that crystal fractionationists have no difficulty in explaining how differentiation has operated in the depths of the earth, but when confronted by a body that has differentiated on an enormous scale before their eyes, so to speak, they are at a loss. In Mark Twain's story, *A Connecticut Yankee in King Arthur's Court*, the magician describes to his audience what eminent personages in distant lands are doing, but is unable to tell what the Yankee is doing with his hand held behind his back.

In spite of these great difficulties there is no doubt that many phenomena are in accord with the theory. As Dr. Bowen has said at times in his writings, "The results are exactly what the theory requires." Is this sufficient proof that the postulated process was the one that was operative? Let us see.

In order to be wholly convincing, the evidence should not only be in accord with the theory but should rule out other possibilities. Perhaps this is too much to require. We shall, therefore, take upon ourselves the onus of disproof and try to show that in certain important instances rocks that are in accord with the requirements of the theory have been formed in a different manner.

I have been shown specimens of rock from the roof of the Bushveld lopolith, collected by Wright, that would generally be regarded as granites, and I am told that rocks of similar character occur in great thickness over hundreds of square miles of the Bushveld. In composition and in relation to underlying rocks they conform, to the

[5] *Platinum Deposits and Mines of South Africa* (Edinburgh and London: Oliver & Boyd, 1929), p. 47.

best of my knowledge, to all the requirements that the theory of crystal fractionation asks for; yet they were not formed in that way. In a strict sense they are not igneous rocks at all, but were produced by the granitization of sedimentary quartzites by emanations. Similar granitization of quartzites and schists on a large scale elsewhere has been described, and the process is not uncommon. I wonder if some of the rocks so formed may not have been regarded as excellent examples of crystal fractionation. What criteria does the theory offer to prevent such a claim from being made in perfect good faith? Or, to put a broader question, have any criteria been given that would prevent almost any rock of igneous aspect, however formed, from being claimed for this process?

We shall take another example, that supplied by the work of Adams and Barlow in the Haliburton and Bancroft areas of Ontario.[6] This work has been acclaimed as a masterpiece of geological research, but the bearing of the author's results on theories of magmatic differentiation seems not to have been generally perceived.

In this area great batholiths of granite invaded limestones, deep in the crust, and brought about what might be termed "hypermetamorphism." This transformation occurred during the active invasion of the magma. The emanations must therefore have been given off by a molten magma and could hardly have been other than gaseous. Great masses of limestone were transformed into silicates and related minerals. Analyses were made, and the results were found to have close resemblance to those of "rocks which are commonly known as diorites, gabbros, basalts, diabases, and essexites."

The following passage is further quoted from Adams and Barlow:

> A remarkable fact in connection with the alteration is that the granite is an acid variety, containing a very small amount of biotite as its only bisilicate, and has, notwithstanding this fact, transfused into the limestone not only silica, alumina, and alkalis, which might be expected, but also large amounts of magnesia and iron. The limestone evidently fixed certain exhalations of the granite magma in relatively greater abundance than others, exerting a species of selective action.[7]

[6] F. D. Adams and A. E. Barlow, "Geology of the Haliburton and Bancroft Areas," *Can. Dept. Mines, Geol. Surv. Branch Mem. 6* (1910).

[7] *Ibid.*, p. 110.

Here, then, are masses of basic rocks of igneous aspect, formed by processes in which crystal fractionation played no part. If the relations had been so concealed that the true origin of the rocks could not have been determined, might they not well be regarded as examples of crystal fractionation?

We should also give attention to the effect of the exhalations upon the composition of the magma itself. Evidently certain constituents departed from it and its composition was changed thereby; this was, of course, a process of differentiation. Likewise, there might very well be, in this case and in others, a differential transfer of material by vapors from lower to upper levels within the magma body itself, effecting differences of composition at various levels. The theory of crystal fractionation does not recognize such processes, and denies that transportation by vapors is of sufficient importance to have much effect. That seems to disregard plain evidence. In addition to the phenomena described by Adams and Barlow, instances of the large-scale conversion of limestones into silicate rocks by magmas have been described by Lacroix, Kemp, and many others. Actually, the direct evidences of such processes are of much greater magnitude than anything that can be pointed to as having surely resulted from crystal fractionation.

In effect, the theory does not admit that any portion of a magma can be split off, nor that variations of composition from one part of the body to another can be brought about, except through the agency of crystallization. I do not believe that this can be effectively maintained.

I shall take up now another set of phenomena. These are of such character that, if everything else appeared to be in accord with the theory of fractional crystallization, I believe that they alone would cast grave doubts upon its actuality as a predominant process. They imply characteristics of magmas fundamentally at variance with those postulated by the theory.

For many years a controversy has been carried on as to whether or to what extent magmas are capable of assimilating solid rock. It has been held by many that what appeared to be assimilation was in reality only disintegration and working over according to the reaction principle. The great obstacle to assimilation, as was repeatedly

pointed out and probably generally admitted, was the supposed inadequacy of the heat supply. Theory and observation were in conflict, and theory seemed to get the better of it.

I have been specially interested in this problem for many years, and, having it much in mind, have been fortunate enough to come upon two instances in the field in which the evidence of great assimilation is so plain that, I believe, any geologist of field experience seeing it would be convinced. What adds to the force of the demon-

FIG. 2.—Banded pumices from Valley of Ten Thousand Smokes. ×0.5

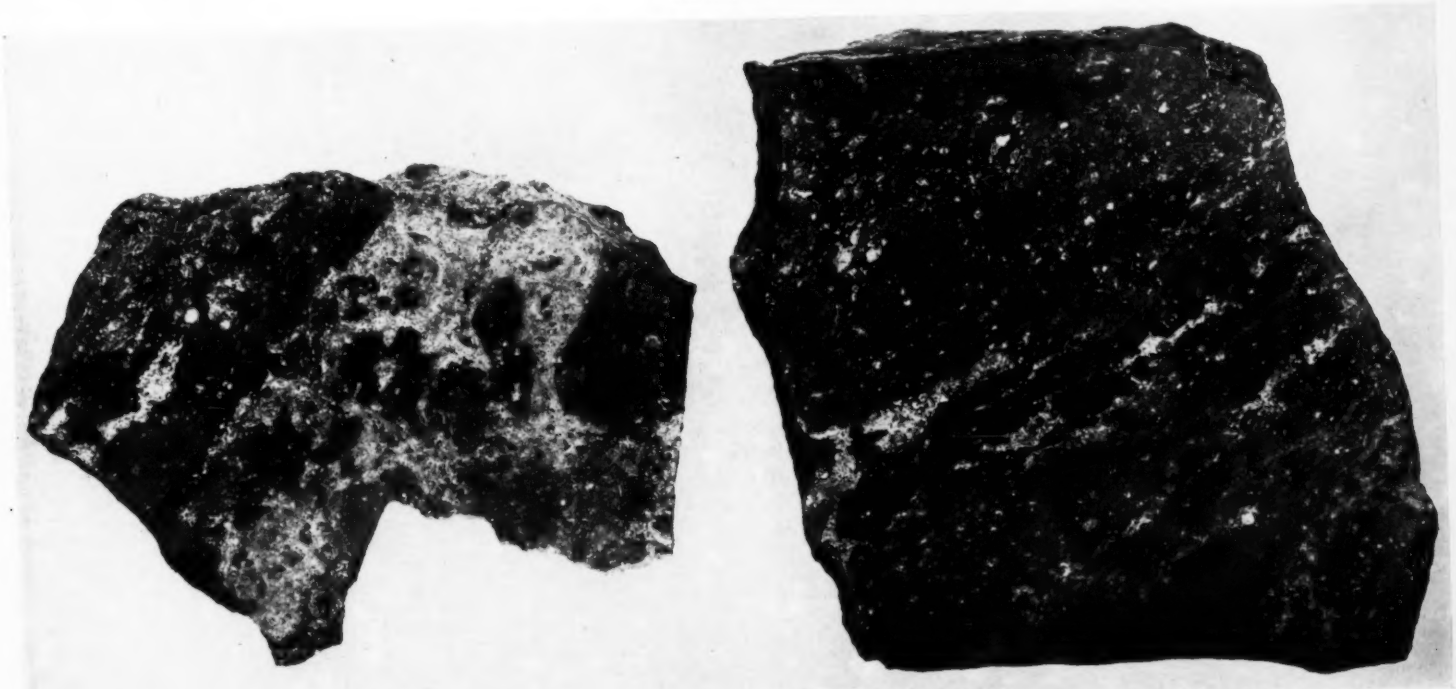
stration is that in both instances the assimilating magma was a surface extrusion of rhyolite and the assimilated rock was very basic.

The first example was in the Katmai region in Alaska. As someone has remarked to me, Alaska is too far away for one to visit. Probably many geologists have believed that there must be some other explanation than the one I have given, but such specimens as those shown in Figure 2 hardly admit of any other interpretation. The white portions are the new rhyolite; the dark represent basic inclusions, almost completely melted and drawn out into bands. Is there any doubt as to what has happened to the basic material? Similar specimens are to be seen by thousands—probably millions—in the Valley of Ten Thousand Smokes. In the adjacent dome of

PLATE I

SOLUTION OF BASALT IN RHYOLITE, GARDINER RIVER, YELLOWSTONE PARK. ×1.35

PLATE II



SOLUTION OF BASALT IN RHYOLITE, GARDINER RIVER, YELLOWSTONE PARK. THE RIGHT-HAND SPECIMEN SHOWS ESPECIALLY THE INTIMATE MANNER IN WHICH THE RHYOLITE FREQUENTLY PENETRATES THE BASALT AND "SOAKS" THROUGH ITS MASS. ×1.35

Novarupta, rock in place is found, and all the transitional stages may be observed by which large and small blocks of basic xenoliths have been acted upon by the rhyolitic magma to form the bands.

The other instance is on Gardiner River in Yellowstone Park. Years ago Iddings apparently came upon a small outlier of this area, and described it without hesitation as an instance of assimilation. In the main area the relations are incomparably more striking. Fortunately the area is now only ten or fifteen minutes' walk from a main highway. In 1934 I showed it to a number of eminent geologists. Among them were Professor and Mrs. Knopf, Professor Larsen and two graduate students, Professor Chamberlin, Professor Bucher, Professor Rouse, and Professor Morris. None of them expressed any doubt or offered an explanation different from mine. Their only query was: "How can it be accounted for?"

From the manner in which the rhyolite flow has penetrated and dissolved the basalt (Pls. I and II), it must have had available energy reserves greatly surpassing those represented by ordinary specific heats. Even apart from the effects of assimilation, the chilling effect of the mere contact of the basalt upon the myriad minute tongues of rhyolite that penetrated the solid rock for many feet would seem to have been much more than sufficient to freeze the rhyolite, if it had possessed only the heat energy ordinarily attributed to it. While I was working on the exposures and had them actually before me, the effects seemed almost incredible.

I have published a tentative explanation of what I have supposed to be the source of the heat in such instances.[8] I am not disposed to insist upon this as a complete or final explanation, but what I do insist upon is that the rhyolite magma was able to do things that the theory of crystal fractionation declares impossible. I want to emphasize that, and I will repeat it: The rhyolite magma was able to do things that the theory of crystal fractionation declares impossible.[9]

[8] C. N. Fenner in *Ann. Rept. of the Director of the Geophysical Laboratory for 1934*, p. 63.

[9] Bowen (*The Evolution of the Igneous Rocks* [1928], p. 198) discusses some of the necessary consequences of the theory of crystal fractionation, and states in the following terms his disbelief in the ability of acid magmas to dissolve basic inclusions: "These remarks are tantamount to the statement that saturated granitic magma cannot dis-

According to this theory, a rhyolitic or granitic magma is necessarily the coolest of the evolution series, and is one from which basic ingredients have been frozen out. It has no heat reserves by which basic material may be taken back into solution. This is the central idea of the evolution of rhyolitic or granitic magmas from basic. Take that away and there is little left. In actual fact, however, we find that a silicic magma is a fundamentally different entity. Its behavior not only seems to controvert the theory of crystal fractionation, but goes much beyond that. The evidence that magmas possess these great and unsuspected reserves of latent energy gives a new basis for the consideration of many earth processes. That, I think, is of much greater significance to geology than is its bearing upon the theory of crystal fractionation.

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solve inclusions of more basic rocks"; and further: "Whatever origin one may assign to a granitic magma . . . . there seems no escape from the conclusion that it will normally be saturated. The normal effect of granite upon more basic inclusions should therefore be such as has been outlined above." In a number of other places in Bowen's book, between pp. 184 and 221, in the discussions on the effects of assimilation, similar explicit statements are made.

# SIERRA NEVADA PLUTON AND CRUSTAL MOVEMENT

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ABSTRACT

Although there is abundant evidence in the Sierra Nevada that part of the magma of the pluton was guided by the northwest direction of cleavage and folding, it is thought that the primary controls were oriented nearly north-south. They guided the volcanic eruptions of Tertiary and Quaternary time as well as the intrusions of the more remote past.

From this it is inferred that the north-south directions are zones of tension; they occur en echelon, and in barbed relation to the regional northwest trend. They may owe their origin to horizontal drift, in which the Sierra block has shifted northward in relation to the eastern blocks.

A survey of the information concerning crustal motion in surrounding regions seems to demonstrate the existence of such drift. It is suggested that the growth of the pluton was directly related to crustal motion, which generated the magma and controlled its emplacement.

The geologically young uplifts and volcanism may indicate the emplacement at depth of another pluton.

## INTRODUCTION

In an earlier paper the structural features of some intrusions and their wall rocks on the eastern slope of the Sierra Nevada were briefly described.[1] It was pointed out that, although the intrusions were forcefully emplaced, the magma was guided by directions or zones of structural weakness. These structural controls were directions of shear, and they usually coincided with the regional northwest trend of cleavage and folding. The modifications imposed upon the regional trends, as a result of forceful emplacement, were clearly somewhat younger than the intense compression which produced the cleavage. It was not possible, however, to decide whether the intrusions rose toward the end of the folding or at some later time.

As a result of more extensive field work and several helpful discussions with Ernst Cloos, the writer has become convinced of another, and possibly much more fundamental, structural control in the Sierra Nevada. This control seems definitely to relate the rise of magma to crustal movement and may disclose the relation of igneous activity to mountain-folding.

[1] E. B. Mayo, "Some Intrusions and Their Wall Rocks in the Sierra Nevada," *Jour. Geol.*, Vol. XLIII (1935), pp. 673–89.

The purpose of the present paper is to combine new data with some earlier results from the Sierra Nevada in order to define this fundamental control and to trace its influence on the pattern of younger volcanism and block faulting. The origin of the control, and its possible significance to the problem of igneous intrusion, will then be discussed. A consideration of the general features of the Sierra Nevada pluton and its geological setting may be a helpful prelude.

## THE PLUTON IN GENERAL

The existence of a granitoid core in the Sierra Nevada has been known at least since the days of J. D. Whitney and the California Geological Survey.[2] This core, or pluton, which is thought to have formed in Upper Jurassic time,[3] is about 400 miles long and locally more than 70 miles wide. It has invaded intensely folded metamorphic rocks of Paleozoic and early Mesozoic age. Its general trend is northwest, parallel to the coast.

The mountains resulting from late Jurassic folding and igneous intrusion were reduced to low relief before Oligocene time. This erosion stripped the cover from the pluton. The resulting mature surface was warped upward during late Tertiary; and block faulting, which continued through the Pleistocene into the Recent, has divided the uplift into numerous fault-block mountains, with intervening grabens. The largest of these mountain ranges is the Sierra Nevada itself (Fig. 1), with its great length, long, gentle western slope, and lofty, eastward-facing scarp. The dimensions of this range reflect the size of the pluton, which was uplifted as a unit.

The gigantic massif, which is now exposed at the surface in the Sierra Nevada, is not one single intrusion but a vast group composed of many smaller units of somewhat different ages.[4] The behavior of

[2] *Geol. Surv. Calif.*, Vol. I (1865), pp. 199–450.

[3] F. E. Matthes, "Geologic History of the Yosemite Valley," *U.S. Geol. Surv. Prof. Paper 160* (1930), pp. 22–23.

[4] This fact, which has been known for many years, has been discussed in a number of papers, among which are the following: W. Lindgren, "The Granitic Rocks of Pyramid Peak District, Sierra Nevada, California," *Amer. Jour. Sci.*, Vol. III (1897), pp. 301–14; H. W. Turner, "The Granitic Rocks of the Sierra Nevada," *Jour. Geol.*, Vol. VII (1899), pp. 141–62; A. Knopf and P. Thelen, "Sketch of the Geology of

certain individual intrusions in a rather large area is one of the chief concerns of this paper.

FIG. 1.—Index map of the state of California, showing location of the Sierra Nevada, and the area shown in Figures 2 and 6.

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Mineral King, California," *Univ. Calif. Dept. Geol. Bull. 4* (1905), pp. 227–62; A. Knopf, "A Geologic Reconnaissance of the Inyo Range and Eastern Slope of the Southern Sierra Nevada, California," *U.S. Geol. Surv. Prof. Paper 110* (1918), pp. 61–72; F. C. Calkins, "The Granitic Rocks of the Yosemite Region," *U.S. Geol. Surv. Prof. Paper 160* (1930), pp. 120–29; H. D. Erwin, "Geology and Mineral Resources of Northeastern Madera County, California," *Calif. Jour. Mines and Geol.*, Vol. XXX (1934), p. 19; Ernst Cloos, "Der Sierra Nevada Pluton," *Geol. Rundschau*, Vol. XXII (1931), Heft 6, pp. 372–84, and *Neues Jahrb.* (1936), in press; "Mother Lode and Sierra Nevada Batholith," *Jour. Geol.*, Vol. XLIII (1935), pp. 225–49.

### RECENT FIELD WORK

During the summer of 1935 a structure section some 60 miles long and 5 miles in average width was surveyed across the Sierra,[5] beginning near Round Valley, about 14 miles northwest of Bishop, and extending across Mount Goddard and Kaiser quadrangles to Black Mountain, in the western foothills. For convenience, this is called "San Joaquin section." The earlier mapping along the front of the range was extended southeastward to connect with unpublished work by K. L. Chasey. The last gap is now closed between Ernst Cloos's Yosemite section[6] and Knopf's reconnaissance along the eastern slope of the southern Sierra.[7] The work so combined does not form a compact unit, but it does outline the structural features of a large area.

### YOSEMITE AND SAN JOAQUIN SECTIONS

*In general.*—The eastern portion of the Yosemite section, together with the San Joaquin section and recent field work along the crest and eastern slope of the Sierra, is shown on the map (Fig. 2). The pluton is bordered along the crest of the range by an almost continuous septum of metamorphic rocks. In the downthrown block, to the east of this septum, is a little-known group of intrusions, mostly concealed by younger deposits.

Some more local features shown are as follows: (1) Tuolumne Complex, with nearly north-south axis; (2) Roof Remnant, a tattered shield of metamorphic rocks; (3) Rock Creek Salient, where the crestline septum is broken and bulged northeastward; (4) Pine Ridge Funnel, a nearly circular intrusion, flaring upward.

In the northwestern part of the map appears the chain of wall-rock fragments that separates the pluton into two major groups within the Yosemite section. In the southwestern corner is a thin, shattered septum chosen as the western boundary of the San Joaquin section. Between Pine Ridge Funnel and Rock Creek Salient are

[5] The methods used in this work were those developed by Hans Cloos for the structural study of igneous intrusions. A very complete treatise on these methods by Robert Balk will be published as Memoir 5 of the Geological Society of America.

[6] "Der Sierra Nevada Pluton," *loc. cit.*

[7] "A Geologic Reconnaissance of the Inyo Range and Eastern Slope of the Southern Sierra Nevada, California," *loc. cit.*

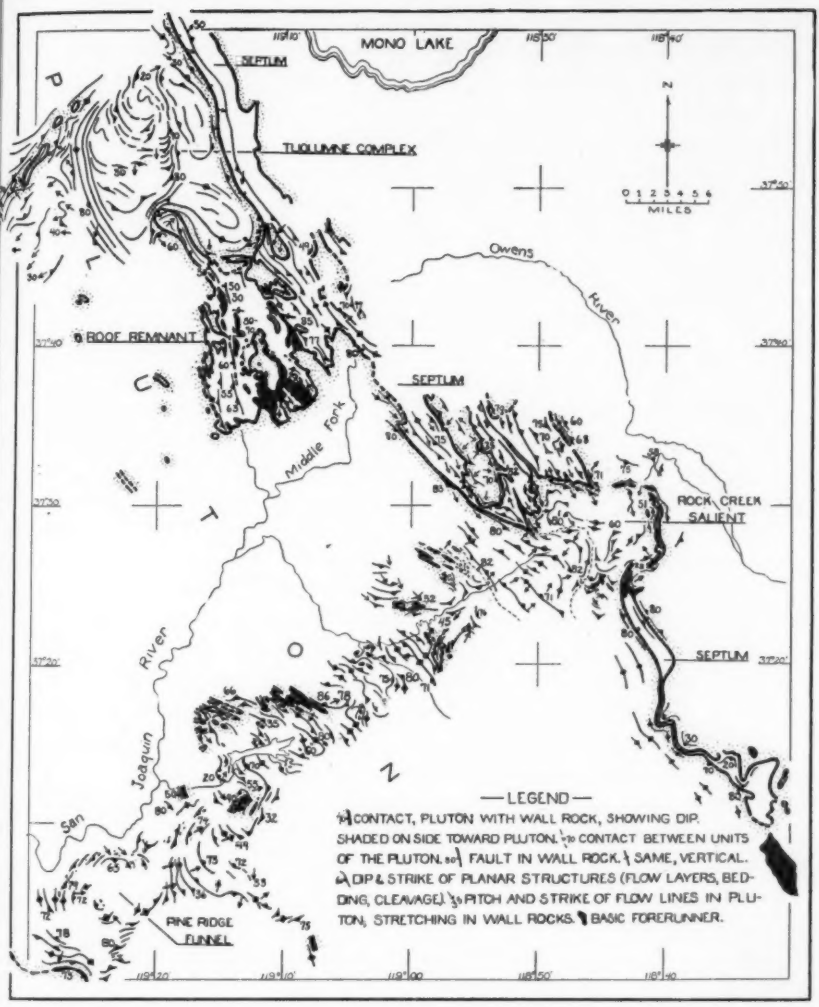
MONO LAKE
SEPTUM
TUOLUMNE COMPLEX
ROOF REMNANT
SEPTUM
Owens River
Middle Fork
ROCK CREEK SALIENT
SEPTUM
San Joaquin River
PINE RIDGE FUNNEL
N
0 1 2 3 4 5 6
MILES
— LEGEND —
CONTACT, PLUTON WITH WALL ROCK, SHOWING DIP. SHADED ON SIDE TOWARD PLUTON. CONTACT BETWEEN UNITS OF THE PLUTON. FAULT IN WALL ROCK. SAME, VERTICAL. DIP & STRIKE OF PLANAR STRUCTURES (FLOW LAYERS, BEDDING, CLEAVAGE). PITCH AND STRIKE OF FLOW LINES IN PLUTON, STRETCHING IN WALL ROCKS. BASIC FORERUNNER.

FIG. 2.—Map of some of the known structures in the pluton and its metamorphosed wall rocks. The structure of Tuolumne Complex is by Ernst Cloos; Roof Remnant is by H. D. Erwin, and the thin septum in the southeastern corner of the map is from unpublished field work by K. L. Chasey.

two or more ill-defined, curving zones of wall-rock fragments. It is thought that these zones represent the shattered remains of septa. If this is the case, then in the San Joaquin section the central massif is divided into at least three composite part-plutons. But this is not all; beyond the western limit shown are two more intrusive groups, separated by nearly half a mile of tightly folded schists. The westernmost of these intrusions is partly covered by sediments in the San Joaquin Valley. In place of two part-plutons, found in the Yosemite section, at least five are developed farther south. Relative ages could not be determined satisfactorily, except in the case of the eastern part-pluton, which is plainly younger than its neighbor on the west.

The increase in the number of part-plutons is paralleled by an increase in width of the entire mass. In the Yosemite section, it is 25–33 miles wide, whereas farther south the width has more than doubled; the igneous core is here more than 70 miles wide. These figures do not include any intrusions beyond the eastern front of the range.

*Internal structure.*—A characteristic of the Yosemite section is the presence of structural domes or schlieren arches.[8] Such structures were not found within the southern section, although in some places the occurrence of relatively gentle dips suggested the lower portions of arches, of which the upper parts had been eroded. With these local exceptions, flow elements in the southern area all pitch or dip very steeply. In order to compare these diverse structural conditions, the eastern portions of both sections are shown in Figure 3.

The Tuolumne Complex is a series of intrusions, one within the other. Half-Dome quartz-monzonite, the oldest member, forms the outer shell of the complex. It was intruded by Cathedral Peak granite—a coarse rock, characterized by huge phenocrysts of potash feldspar. In the center of the complex is an aplitic mass, the Johnson granite porphyry.[9] As shown in Figure 3, the flow structure of the complex defines a number of nested, broad arches through which the central intrusion breaks discordantly.

Rock Creek Salient is the southern analogue of Tuolumne Com-

[8] Ernst Cloos, "Der Sierra Nevada Pluton," *loc. cit.*, p. 377.

[9] See F. C. Calkin's map, *U.S. Geol. Surv. Prof. Paper 160* (1930), p. 122.

plex. Here the Johnson granite-porphyry may be absent as such, but its place is taken by a nearly structureless, silicic, central variant of Cathedral Peak granite. Around the latter, which is again characterized by huge feldspar phenocrysts, is a partial girdle of quartz-monzonite and granodiorite. The internal structure, however, differs from that of the Tuolumne mass. There are no schlieren arches, no structural domes. Instead, the attitudes of the flow elements are everywhere steep.

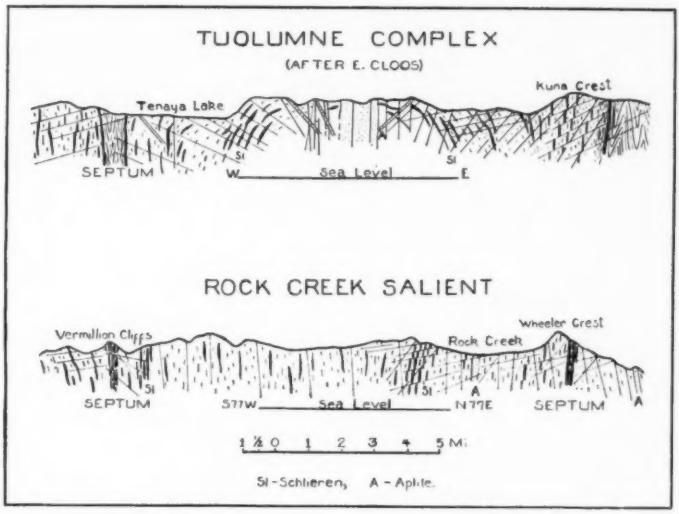


FIG. 3.—Structure sections through Tuolumne Complex and Rock Creek Salient. In Tuolumne Complex, schlieren domes and a fan of cross-joints suggest a high level in the intrusive body. A lower level is indicated by the vertical structures at Rock Creek Salient. In both sections gently dipping marginal thrusts are more abundant along the eastern border.

Experimental intrusions have been made to develop arching flow structures near their roofs, while the traces of flow were nearly vertical in the lower portions.[10] Accordingly, the arching structures in

[10] W. Riedel, "Das Aufquellen geologischer Schmelzmassen als plastischer Formänderungsvorgang," *Neues Jahrb. f. Min., etc.*, Vol. LXII, Abt. B (1929), pp. 151–70; H. Cloos and E. Cloos, "Das Strömungsbild der Wolkenburg im Siebengebirge," *Zeit. f. Vulkanologie*, Vol. XI (1927), p. 95.

the Tuolumne Complex must indicate a rather high level, near a broad roof. The aplitic mass at the center may have broken into, or even through, the cover. The essentially vertical structures at Rock Creek Salient doubtless indicate a considerably lower level in the intrusive body, which as a whole must pitch northward. Pine Ridge Funnel, near the western end of the San Joaquin section, may represent a still lower level; its downward convergence and nearly circular shape suggest a pipelike feeding channel.

This evidence of northward pitch is of special interest in view of Ernst Cloos's theory of the relation of ore values in the Mother Lode zone to the attitude of the pluton.[11] So far the evidence from eastern and western borders of the massif is in complete agreement. It would be interesting to study more sections, located farther south, to see whether this relation may hold throughout the entire length of the Sierra.

*The surface pattern of flow layers.*—The arrangement of flow layers within the massif is obviously very complicated, as is shown in the Yosemite and San Joaquin sections. The pattern of the flow layers resembles somewhat that formed by bands of foam on water undergoing slow but turbulent motion. Local trends appear to be entirely independent of any regional direction.

Were the pluton a single, huge unit, it seems that such a flow pattern would mean nearly perfect freedom of motion. Under this condition the magma might be compared to a pot of boiling porridge or to fluent lava seething in an open crater. During the progress of field work it was tempting to regard this complicated flow pattern as evidence that the magma had welled up under its own power, perhaps long after the time of folding.

The pluton, however, is a composite of many small intrusions and was never a vast sea of molten magma. Each single unit has its own flow structure, which reflects the local conditions under which a particular increment of the magma was emplaced. Widespread turbulence within a single vast reservoir is absolutely precluded in this instance. It is the presence of many individual and often transgressing patterns that complicates the picture. The writer has already presented reasons for thinking that lateral compression could

[11] "Mother Lode and Sierra Nevada Batholith," *loc. cit.*

not have been intense, at the present level of exposures, during the igneous invasion,[12] but even so the igneous activity may not have been completely separated in time from the folding. The reason for this statement lies in an anomalous behavior of certain intrusions and intrusive groups.

### EVIDENCE OF A NEARLY NORTH-SOUTH STRUCTURAL CONTROL

The evidence of a structural control, quite distinct from the shears that parallel or nearly parallel the northwest regional direction, is plainest near the eastern margin of the pluton. It has been worked out most completely at Tuolumne Complex and Rock Creek Salient. From the characteristic features at these localities, it is possible to infer with some confidence the presence of the same control at other places.

*Tuolumne Complex and Rock Creek Salient.*—From the northwestern corner of the Roof Remnant, a long, narrow zone of wall rock projects in the direction of the regional trend. At the far end of this zone, cleavage in these metamorphosed rocks is sharply deflected, as though to avoid some obstacle that does not accord with the regional structure. This obstacle is the central portion of Tuolumne Complex, which trends slightly east of north. Obviously, this massif was not guided by any regional attitude of cleavage or shears. Instead, it has sought out and enlarged a nearly north-south, discordant zone of weakness.

Just west of Rock Creek Salient the Cathedral Peak granite forms a long, narrow intrusion, oriented about N. 40° W., parallel to the regional trend. Opposite the salient this mass abruptly bulges northward or northeastward. The intrusions which preceded Cathedral Peak granite at this locality likewise bulged northward to an even more remarkable degree. As a result the crestline septum, which has generally a rather uniform northwest trend, is locally bent far out of its course and partially disrupted. Here again, some nearly north-south control has locally favored the rise of magma.

*Coyote Ridge and vicinity.*—Of the features shown on Knopf's reconnaissance map,[13] those south and east of Coyote Ridge are most interesting in relation to the present discussion. Coyote Ridge lies

[12] Mayo, *op. cit.*

[13] *U.S. Geol. Surv. Prof. Paper 110* (1918).

at the southeastern edge of Figure 2, and the areal relations near there are shown in Figure 4. The crestline septum, entering the area from either north or south, is bent abruptly eastward and squeezed out between adjacent massifs.

It seems as though this septum may once have described a great curve, similar to, but larger than, the one at Rock Creek Salient.

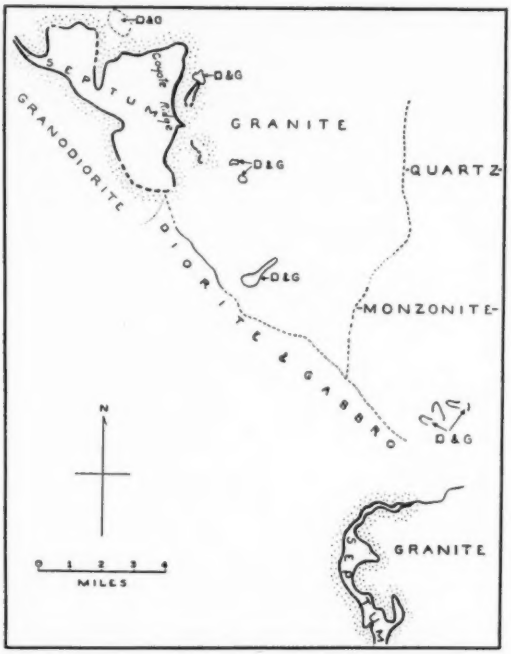


FIG. 4.—Areal geologic map (after A. Knopf, with additions from the work of K. L. Chasey) of the area adjoining Coyote Ridge in Inyo County. Apparently a northeastward bulge in the septum has been torn asunder by intrusions of quartz monzonite and granite which followed a north-south direction. Fragments of the septum, and of diorite and gabbro, float in these discordant intrusions.

The conditions at the salient, however, have been carried to a much more advanced stage southeast of Coyote Ridge: The curving portion of the septum has been disrupted, and a breach some 12 miles wide exists between the broken ends. This breach was filled mainly

by quartz-monzonite and granite, in which pieces of the wall rock float as isolated fragments. Granite, youngest of the intrusions, is an enormous mass, the northern end of which forms the Tungsten Hills, west of Bishop.[14] The southern end of this great intrusion traverses the septum and projects into the complex east of Coyote Ridge. The last major event in the intrusion history was the emplacement of this granite, which clearly follows a north-south direction.

There exists one more locality near the eastern margin of the pluton, where a north-south control is suspected. Before this locality is discussed, it is necessary to consider the joint trends of the region.

*Joint patterns.*—Two general directions, one N. 30°–50° E., the other N. 30°–50° W., are favored by most of the regional joints everywhere in the Sierra. The northwest direction roughly parallels the regional trend of cleavage and folding; the other traverses it at right angles. Usually, the northeast direction is more strongly favored, and locally it controls the topography, as shown by the drainage pattern on the topographic map of Yosemite National Park. The sheer, northwestern face of Half-Dome is controlled by the northeast joints, and the same is true of many canyons in the northern part of the park.

Locally, a third joint direction (N. 0°–25° E.) is dominant as, for instance, within the Tuolumne Complex.[15] The joint patterns at and near Rock Creek Salient are shown on Figure 5. West and northwest of the salient, as indicated by the joint diagrams, the usual northwest and northeast joints prevail. As the salient is approached these regional sets become suppressed, and a new trend, N. 10°–20° E., dominates. The upper course of Rock Creek follows this direction.

To an observer on one of the lofty summits near the head of Rock Creek the oblique rays of the late-afternoon sun disclose many

[14] Knopf, "Tungsten Deposits of Northwestern Inyo County, California," *U.S. Geol. Surv. Bull. 640-L* (1927), pp. 229–49; F. L. Hess and E. S. Larsen, "Contact-Metamorphic Tungsten Deposits of the United States," *U.S. Geol. Surv. Bull. 725-D* (1921), pp. 268–74.

[15] Ernst Cloos, personal communication.

thousands of northward-trending joints in the granite mass east of Coyote Ridge. Thus it is clear that, wherever magma was guided by nearly north-south zones of weakness, swarms of steeply dipping joints follow the same direction.

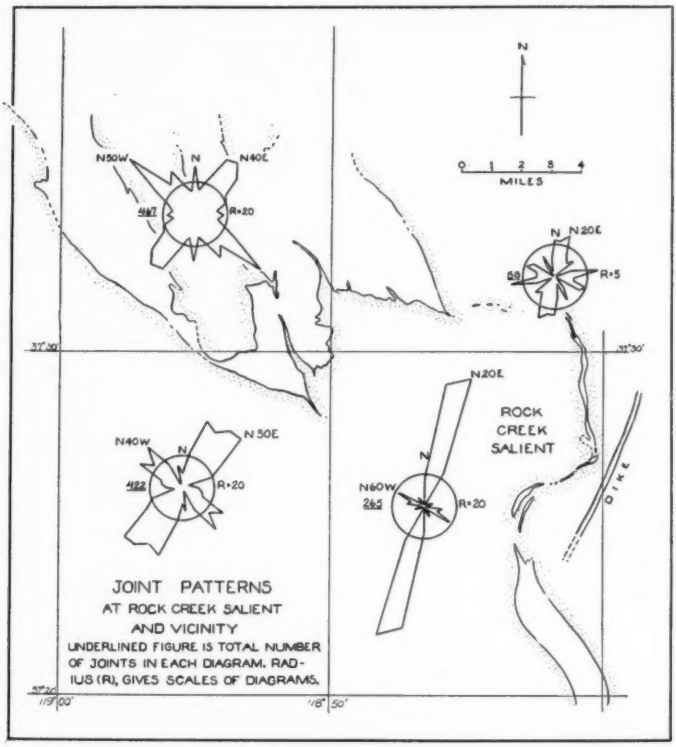


FIG. 5.—Joint patterns near Rock Creek Salient. The regional northwest and northeast joint directions show plainly in the two left-hand diagrams. A direction near N. 20° E. appears in the salient, and farther east a huge alaskite dike follows a similar trend.

*Middle Fork.*—Since the present streams are mostly joint-guided, it should be possible by using the drainage patterns to pick out other places where nearly north-south joints prevail. Such a locality seems

to exist on Middle Fork of San Joaquin River, where that stream follows a northward course. Here, again, the local departure of joint trends from the regional directions is accompanied by a corresponding anomalous behavior of intrusions: Whereas most of the intrusive lobes that penetrate the Roof Remnant from the south are elongated parallel to the regional northwest direction, the intrusion on Middle Fork plainly shows a tendency to break into the metamorphics in a north-south direction.

From these observations it is evident that at the time the intrusions were emplaced discordant zones of weakness oriented slightly east of north occurred at intervals along the eastern border of the pluton. The former existence of these zones is now recorded by the orientation of intrusive axes (Tuolumne Complex, Middle Fork, granite east of Coyote Ridge); by the direction in which intrusions bulge (Rock Creek Salient); and by local departures from the regional trends of joints. This direction is oriented to the northwest regional trend as the barbs of a feather are oriented to the shaft. At the localities mentioned above it has favored and guided the rise of magma.

Not enough field work has yet been done to enable one to state to what extent north-south structures occur within the pluton. It is thought, however, that the eastern margin is not unique in possessing such trends. Some facts now at hand plainly suggest the same control within the central massif.

On Cloos's Yosemite section flow patterns both east and west of Sentinel Dome seem to indicate part-plutons with north-south axes. The western contact of the Roof Remnant (Fig. 2) may owe its northward trend to forceful emplacement of an intrusion that followed this direction.

On San Joaquin section it is thought that the conditions described at Rock Creek Salient may have been repeated west of each of the curving zones of wall-rock fragments. These arcuate zones are remarkably similar in plan to the curving septum at Rock Creek. Assuming that the intrusions on this southern section are progressively younger from west to east, as E. Cloos found to be the case on the Yosemite section, then the same structural condition must have existed repeatedly as the locus of intrusion migrated eastward.

Early stages are preserved in the arcuate zones of wall-rock fragments. The last step is recorded at Rock Creek Salient.

### YOUNG FAULT PATTERN AND ASSOCIATED VOLCANISM

The pattern of Tertiary and Quaternary faults and fissures near the eastern front of the Sierra is shown in Figure 6. Most of the fractures shown are faults; gaping fissures that usually trend northward occur locally. These faults and fissures are represented by solid lines wherever the evidence (usually physiographic) seemed conclusive in the field. Broken lines indicate faults deduced from the topographic maps or for which the field evidence was only suggestive. The vertical component of motion along some of the faults is responsible for most of the relief in this region.

The structural "grain"—that is, the regional trend of cleavage and folding—is followed by faults which determine the general trend of the mountain front. The rather close dependence of these young faults upon the ancient structure is best shown at Rock Creek Salient, where the bounding faults swing around the curving septum and bulging intrusions as wood splits around a knot.

If the northwest direction were the only control, the Sierra front should follow a more or less regular course, marred by smooth curves and occasional abrupt bends like the one at Rock Creek. Actually, the mountain front is characterized by local zigzags, projecting buttresses, and broad re-entrants. Some of the buttresses are plainly blocked out by northeast faults.[16] Thus there are northwest and northeast faults parallel with the directions of regional jointing.

If the northeast and northwest regional joint sets have their counterparts in the pattern of the younger faults, the nearly north-south joint direction is no less favored. Swarms of north-south faults and fissures locally impinge against the Sierra scarp. Wherever this takes place the scarp recedes and a broad re-entrant is formed. These geologically young faults and fissures seem to correspond approximately in direction with the structures that guided discordant intrusions in Upper Jurassic time. It should be noted,

[16] The intersection of northwest and northeast faults locally gives the scarp a zigzag course. A lunar example, the jagged escarpment of the Altai Mountains of the moon, has been described by E. H. L. Schwarz, in *Jour. Geol.*, Vol. XXXVI (1928), p. 105.

however, that whereas the earlier structures are usually oriented 10°–20° east of north, the corresponding young faults trend more

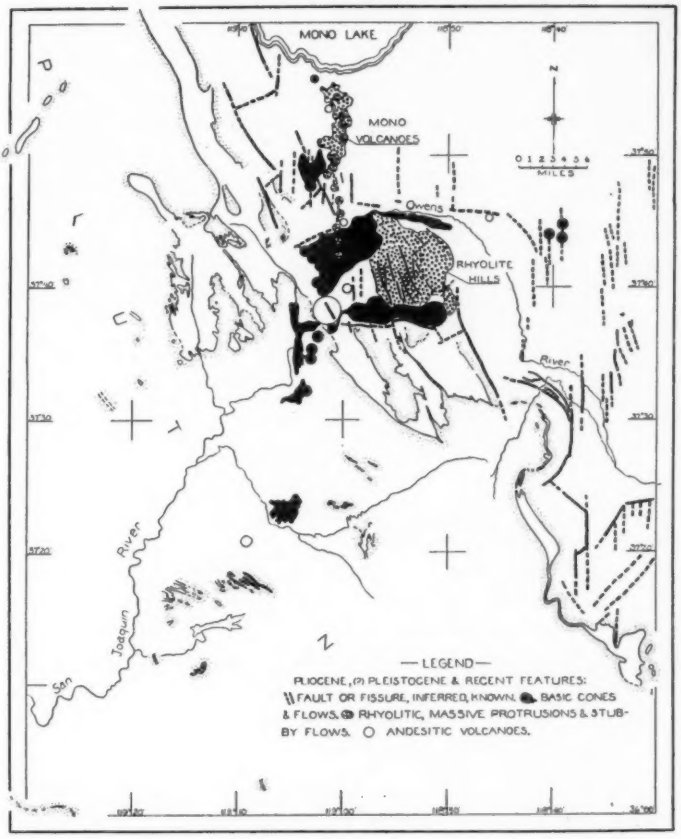


FIG. 6.—The Tertiary and Quaternary fracture pattern in the area of Figure 2. Northwest faults control the general trend of the Sierra; northeast ones produce local irregularities in the scarp. North-south faults and fissures guide most of the volcanic eruptions and produce broad re-entrants in the Sierra front. The Rhyolite Hills occupy an intensely fissured area of subsidence in one of the re-entrants.

nearly due north. This suggests that the earlier structures may have been rotated clockwise since they were formed.

It is not surprising that the younger magmas of the region have been guided by north-south structures, just as were their predecessors of the Upper Jurassic. The young volcanoes and lava flows known to the writer are shown in Figure 6. Many more could have been indicated if the data were at hand. Wherever these eruptions were basic, they built small cinder cones and spread far beyond them as thin flows; wherever acid and viscous, they formed massive protrusions and short, stubby flows.

Among the acidic eruptions the most remarkable are the Mono volcanoes, a chain of lofty domes extending southward from Mono Lake. On the assumption that the magma has risen along faults, it seems at first that these volcanoes should follow the northwest trend of the Sierra front. Their independent northward course results from the discordant, north-south control.

Slightly west of the Mono volcanoes, and roughly paralleling them, is a chain of basic cinder cones with far-spread lava flows. In places, groups of these cones are alined to the northwest direction, but this orientation is a very local feature. The general trend of the basic eruptions is north-south. Since this chain of volcanoes extends far within the pluton, some of the north-south fractures must have bitten deeply into the Sierra block.

About 18 miles south of Mono Lake a great re-entrant exists in the Sierra front. The Mono volcanoes trend into this re-entrant, and many other meridional structures strike into it. It appears that here a block of the crust, weakened by many fissures, has subsided below the general level. This block is encircled by faults, and tremendous volumes of lava and pyroclastic materials have been expelled along its northern and eastern sides.[17] Protrusions and stubby flows of rhyolite (Rhyolite Hills) have issued upon the intensely fractured floor of the re-entrant. Since the plan of these hills and encircling faults resembles the structural plan of Rock Creek Salient, it is thought that the Rhyolite Hills and their immediate surroundings may be the surface expression of a deep-seated condition like the one now exposed at Rock Creek.

[17] These volcanic deposits are not shown on the map. The fault pattern and volcanic history of the country north and east of Owens River is being studied in detail by Mr. C. M. Gilbert.

At least twice in the history of the Sierra (in Upper Jurassic time and during the period of block faulting) north-south structures have guided the rise of magma, so that the resulting intrusions or extrusions cross the regional trend of cleavage and folding. This observation, together with the fact that the north-south direction is locally followed by open fissures, indicates east-west tension. What has caused such tension, and why did it produce en echelon structures, showing a barbed relation to the northwest regional trend? Also, why have the older structures in most cases apparently been rotated 10°–20° east of north, while the younger ones usually trend due north?

### EN ECHELON TENSION FRACTURES

The origin of en echelon tension fractures has been the subject of many papers, and a number of theories have been proposed to account for such structures. The views that have come to the writer's attention may be grouped as follows: (1) compaction of sediments over buried ridges;[18] (2) differential uplift or subsidence, with[19] or without[20] torsion; and (3) shear.[21] Of the explanations suggested, (1) obviously does not apply to the problem at hand. In the Sierra region, where very marked differential uplifts have taken place, the suggestions of Foley[22] and Sherrill[23] cannot be put aside. Their explanations doubtless find some applications in this area,

[18] Theodore A. Link, "En Echelon Tension Fissures and Faults," *Amer. Assoc. Petrol. Geol. Bull. 13* (1929), pp. 627–43 (with discussion).

[19] R. E. Sherrill, "Origin of the En Echelon Faults in North-Central Oklahoma," *Amer. Assoc. Petrol. Geol. Bull. 13* (1929), pp. 31–37.

[20] Lyndon L. Foley, "Mechanics of the Balcones and Mexia Faulting," *Amer. Assoc. Petrol. Geol. Bull. 10* (1926), pp. 1261–69.

[21] A. E. Fath, "The Origin of the Faults, Anticlines, and Buried 'Granite Ridge' of the Northern Part of the Mid-continent Oil and Gas Field," *U.S. Geol. Surv. Prof. Paper 128-C* (1920), pp. 75–84; Foley, "The Origin of the Faults in Creek and Osage Counties, Oklahoma," *Amer. Assoc. Petrol. Geol. Bull. 10* (1926), pp. 293–303; William Kramer, "En Echelon Faults in Oklahoma," *Amer. Assoc. Petrol. Geol. Bull. 18* (1934), pp. 243–50; R. T. Chamberlin, "A Peculiar Belt of Oblique Faulting," *Jour. Geol.*, Vol. XXVIII (1919), pp. 602–13; Hans Cloos, "Bau und Bewegung der Gebirge in Nordamerika, Skandinavien, und Mitteleuropa," *Fortschr. d. Geol. u. Paläont*, Band VII, Heft 21 (1928), pp. 241–327.

[22] "Mechanics of the Balcones and Mexia Faulting," *loc. cit.*

[23] *Op. cit.*

but, as Hans Cloos pointed out in 1928, the regional fracture pattern may be primarily the result of shear.[24]

The fact that en echelon tension fractures develop in barbed relation to shears has been known for a long time. The cracks formed in 1906 by horizontal movement along a part of the San Andreas rift were so related to the general trend of the rift. The mechanical conditions of their formation have been explained by H. F. Reid.[25] The same type of structure has been reproduced experimentally by W. J. Mead,[26] W. Riedel,[27] and Ernst Cloos.[28] They exist in nature as the well-known marginal crevasses, recently analyzed mathematically by M. Lagally.[29] They occur along intrusive boundaries, faults, and thrusts, where it has been proposed to call them "feather joints." On a vaster scale they may exist as regional structures. The Christiania region of Norway is said to be a system of N.–NNW.-trending "Fiederspalten."[30] The zone, in which these fractures lie, trends east of north. Huge rhombenporphyry dikes, chains of essexite intrusions, and some large laurvikite and granite massifs follow the tension fractures.

Whether developed on a regional scale or present as tiny gash veins, the acute angle made by these fractures with the surface of motion always opens in the direction from which the motion came. Applying this principle to the eastern front of the Sierra Nevada, it would seem that the Sierra block and the country west of it has moved northward relatively to the Great Basin. If such motion had been active throughout late Tertiary and Quaternary time, it would probably have produced most of the structures shown in Figure 6.

[24] *Op. cit.*, pp. 281–92.

[25] "Report of the State Earthquake Investigation Commission," *Carnegie Inst. Wash. Publ. 87*, Vol. II (1910), pp. 33–35.

[26] "Notes on the Mechanics of Geologic Structures," *Jour. Geol*, Vol. XXVIII (1920), pp. 505–23.

[27] "Zur Mechanik geologischer Brucherscheinungen," *Centr.-Bl. f. Min., etc.*, Abt. B, No. 8, pp. 354–68.

[28] " 'Feather Joints' as Indicators of the Direction of Movements on Faults, Thrusts, Joints and Magmatic Contacts," *Proc. Natl. Acad. Sci.*, Vol. XVIII (1932), pp. 387–95.

[29] "Versuch einer Theorie der Spaltenbildung in Gletchern," *Zeit. f. Gletcherkunde*, Band XVII, Heft 4/5 (1929), pp. 285–301 (esp. pp. 293–301).

[30] Hans Cloos, *op. cit.*, pp. 294–301.

Similar motion in Upper Jurassic time would account for discordant intrusions and the peculiar behavior of joints along the eastern margin of the pluton. Rotation, resulting from such motion, may account for the fact that the older tension fractures now trend farther east of north than do most of the younger ones.[31]

Associated with horizontal drift in this region, there has been a prominent vertical motion, which is responsible for the topographic relief. Vertical motion will hardly be questioned; the horizontal drift is difficult to accept. Such horizontal motion, to have produced the features described, must have been regional, and it must have affected the crust to very great depths. It could hardly have failed to leave its impress on the structure of surrounding regions. A review of some salient features of the structure of California and western Nevada may serve to test this idea of horizontal motion.

## REGIONAL HORIZONTAL MOTION

In presenting the evidence it may be best to begin near the coast, where the structure is better known, and where the motion appears to have been most intense. With a few pertinent exceptions, only horizontal motion along steeply inclined shear zones will be considered since this is the type of movement that bears most directly on the present problem.[32]

*San Andreas rift.*—This, the most important shear zone in the coastal region, has a known length of some 600 miles. It extends from the vicinity of Salton Sink, in southeastern California, to Punta Arenas, on the coast north of San Francisco. A sudden slip along a portion of this rift is held responsible for the San Francisco earthquake of 1906.[33] The arrangement of tension cracks in surface soil along the rift, together with actual offsets of roads, fences, and other markers traversed by the rift, indicated a northward movement of

[31] Hans Cloos (*ibid.*, pp. 271–72), experimenting with flowing clay, noted that after a fracture has formed it may be rotated and later traversed by other fractures that appear in the initial position.

[32] A discussion of the several types of motion in this region, including horizontal drift, thrusting, and normal faulting, has been presented by Hans Cloos (*ibid.*, pp. 281–94).

[33] Report of the State Earthquake Investigation Commission, "The California Earthquake of April 18, 1906," *Carnegie Inst. Wash. Publ. 87* (1909–10).

the western block in relation to the eastern one. The greatest horizontal displacement observed was 21 feet, whereas the maximum vertical displacement was only 3 feet.[34] The same relative movement of the two sides of the rift is indicated by active thrust faults in San Benito County,[35] where thrusts west of the rift are directed northward, while those east of it moved southward.

In the region south of San Francisco Bay, the accumulated amount of horizontal movement along the San Andreas Rift is thought to have been in the order of 20 miles.[36] In the desert region of southeastern California, Noble[37] found local evidence that suggested a shift of 24 miles in the same direction. However, he also found some contradictory evidence. There were places where the western block appeared to have moved southward.

Recent soundings off Cape Mendocino indicate a submarine escarpment, "along which the 1,000-fathom contour has been displaced about 40 miles in the proper direction."[38]

*Newport-Beverly shear zone.*—The Long Beach earthquake of 1933 has been referred to motion along this zone. According to Eaton,[39] displacements along the zone have been mostly horizontal and correspond in direction with those usually found along the San Andreas rift.

*Haywards and Sunol faults.*—Vickery[40] has presented evidence of 9–13 miles of horizontal movement along the Sunol Fault in the Livermore region. Here, again, the western block appears to have moved northward. From the offset course of Strawberry Creek in

[34] These figures are taken from a table compiled by B. M. Page (*Jour. Geol.*, Vol. XLIII [1935], p. 705).

[35] P. F. Kerr and H. G. Schenck, 'Active Thrust Faults in San Benito County, California," *Geol. Soc. Amer. Bull. 36* (1925), pp. 465–94.

[36] *Ibid.*, p. 478.

[37] L. F. Noble, "The San Andreas Rift in the Desert Region of Southeastern California," *Carnegie Inst. Wash. Yearbook 31* (1932), pp. 355–63.

[38] Francis P. Shepard, "The Northern Continuation of the San Andreas Fault" (abst.), *Geol. Soc. Amer. Proceedings* (1936), p. 105.

[39] J. E. Eaton, "Long Beach, California, Earthquake of March 10, 1933," *Amer. Assoc. Petrol. Geol. Bull. 17* (1933), pp. 694–712.

[40] Frederick P. Vickery, "The Structural Dynamics of the Livermore Region," *Jour. Geol.*, Vol. XXXIII (1925), pp. 608–28.

the Berkeley Hills, J. P. Buwalda[41] inferred a "rather recent" 1,400-foot northward shift of the western side of Haywards Fault. R. J. Russell[42] found that recent northward movement of the block west of Haywards Fault had offset some of the larger stream courses about 600 feet; smaller streams were offset 50 or 25 feet.

To the horizontal motion reported along the faults mentioned above should be added an unknown amount of similar motion that must have taken place along other northwest-trending fault zones of the coastal area. It seems certain that the total horizontal displacement along steeply dipping shear zones of northwest trend has been several tens of miles, and in spite of local anomalies it appears that the western portion has moved northward in relation to the interior.

*Sierra Fault zone.*—The motion deduced from the fracture pattern along the eastern front of the Sierra Nevada is matched by that which accompanied the Owens Valley earthquake of 1872.[43] The greatest horizontal displacement observed was 12 feet, and the maximum vertical displacement was 23 feet. A vertical component was dominant, but horizontal motion was obviously present, and it acted in the usual direction.

*Western Nevada.*—Gianella and Callaghan[44] have called attention to what may be an important northwest shear zone in western Nevada. This zone separates a western region of dominantly northwest-trending mountain ranges from an eastern region in which the ranges trend north-south or slightly east of north. The Cedar Mountain earthquake of 1932 occurred on a part of this zone. The pattern of fissures formed at the time of the earthquake indicated horizontal movement in the usual direction.

In each case noted above the observed or inferred displacement oc-

[41] "Certain Inferences Regarding the Nature of the Movements along the Hayward Fault Zone" (abstract), *Geol. Soc. Amer. Bull.* 37 (1926), p. 212.

[42] "Recent Horizontal Offsets on the Haywards Fault," *Jour. Geol.*, Vol. XXXIV (1926), pp. 507–11.

[43] W. H. Hobbs, "The Earthquake of 1872 in Owens Valley, California," *Beitr. Geophysik*, Band X (1910), p. 379. This paper has not been seen by the present writer.

[44] Vincent P. Gianella and Eugene Callaghan, "The Earthquake of December 20, 1932, at Cedar Mountain, Nevada, and Its Bearing on the Genesis of Basin Range Structure," *Jour. Geol.*, Vol. XLII (1934), pp. 1–22.

curred along a northwest-trending fault zone. Recently, Page[45] has presented data on fault movements in a zone that trends somewhat east of north. At Pleasant Valley, Nevada, the horizontal motion resulting from the earthquake of 1915 was negligible, while the vertical displacement was 16 feet. This is the result that would be expected if the regional north-south structures are tension fractures resulting from horizontal shear along zones of northwest trend.

From the evidence discussed so far it seems very probable that, during Tertiary and Quaternary time, the entire region between the Great Basin and the Pacific Ocean has undergone considerable horizontal displacement in the direction required to explain the fracture pattern along the eastern front of the Sierra Nevada. Vertical motion is also recognized and accounts for the present relief. Other factors may have co-operated to produce the fracture pattern, but it seems unnecessary to seek a major cause other than horizontal movement.

### RELATION OF HORIZONTAL MOTION TO THE PLUTON

If the north-south fractures and associated volcanism along the Sierra front resulted from horizontal shift in Tertiary and Quaternary time, is it necessary to find some other cause to explain similar structural features in the pluton? The writer feels that it is more logical to assign these similar structures, however different their ages, to the same origin and to assume that horizontal motion was also active in Upper Jurassic time and influenced the growth of the pluton.

Accordingly, this granitoid massif may parallel an ancient zone of shear. Yet a shear zone is likely to be tight and not to favor intrusion. For instance, there appears to be no evidence that intrusions on a large scale accompanied the movement along the San Andreas rift. But if the differential motion, instead of being restricted to a comparatively narrow zone, were so distributed as to favor the development of regional tension fractures,[46] these might favor the rise of magma. Moreover, it may be worth while to consider whether the generation of magma at depth may not be localized along zones

[45] *Op. cit.*

[46] For a discussion of the relation of tension fractures to various types of differential motion, where ice is the fractured medium, see Lagally, *op. cit.*, pp. 293–301.

of tension. As the magma rises, it may seek out, and mold itself to, the regional shear directions; but at many places the zones of tension will control, and the regional trends will be traversed or modified to fit the discordant massifs (Rock Creek Salient, Fig. 2). As each part-pluton solidifies, it should strengthen locally the crust; further fracturing and intrusion will tend to avoid these strengthened localities.[47]

The Sierra Nevada pluton appears to have started its growth in the west, near the Mother Lode zone, and to have developed eastward.[48] As a result of slow, discontinuous growth a vast assemblage of part-plutons accumulated. Many individual intrusions may be oriented nearly north-south, but the group as a whole parallels the shear direction.

It can hardly be doubted that the magma arrived on the scene after folding was essentially completed;[49] but if the arguments in this paper are sound, the emplacement of a massif like the Sierra Nevada pluton is definitely an accompaniment of mountain-building movements. The motion must have continued a long time before intrusions appeared. Apparently, the older rocks had to be intensely folded, and the crust thereby made more rigid, before conditions were favorable to the generation and intrusion of magma. As the crustal motion died away, growth of the pluton ceased.

Long after the pluton had grown to full size, and after erosion had bared its crystalline core, horizontal motion was renewed. The volcanic activity which accompanied this motion was localized along zones of tension. In the northern, lower portion of the Sierra, the volume of materials extruded has been stupendous. Southward the volcanic activity decreases, but there the eroded surface of the pluton has been uplifted to imposing height. Is this volcanism and differential uplift a surface expression of the growth of a second and much younger composite pluton? If so, the vertical component of motion in this region is explained. The upward surge of the younger massif has broken the crust along the eastern border of the old one.

[47] Where this is not the case, nested intrusions (Tuolumne Complex) will form.

[48] Ernst Cloos, "Der Sierra Nevada Pluton," *loc. cit.*, pp. 372–73.

[49] Mayo, *op. cit.*, pp. 687–88; Walter H. Bucher, *The Deformation of the Earth's Crust* (Princeton University Press, 1933), pp. 274–82.

## CONCLUSION

In spite of the vast amount of work that has been done in the Sierra and in certain portions of the surrounding region, the data collected represent only a small fraction of those still to be obtained. Until much more field work has been done the opinions advanced in this paper cannot be proved. Also, these opinions are not original; they were advanced in part by Hans Cloos in 1928 and will be further elaborated in a forthcoming report by Ernst Cloos.[50] The present paper adds appreciably to the store of information bearing upon the problem. As this information accumulates, the idea of the control of magma through tension that results from horizontal shift seems to become more firmly entrenched. The writer hopes that this idea may become generally known and accepted not as a fact but as a valuable working hypothesis. The structurally unknown areas in the Sierra Nevada are still far larger than the known ones. As the field work progresses, and as more evidence accumulates from surrounding regions, the strength or weakness of the hypothesis must surely come to light.

ACKNOWLEDGMENTS.—The writer's structural work in California was made possible through grants-in-aid from the National Research Council, and to this organization he takes pleasure in expressing his sincere thanks. Ernst Cloos, whose structural surveys in the Sierra are now widely known, has been a never failing source of help and inspiration. His kindness is gratefully remembered. It is also a pleasure to recall the assistance and companionship of Dr. John M. Parker III during the survey of the San Joaquin section.

[50] "Der Sierra Nevada Pluton in Californien," *Neues Jahrb., etc.* (1936, final report, in press).

# SEDIMENTARY GNEISSES OF THE SALMON RIVER REGION NEAR SHOUP, IDAHO

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## ABSTRACT

Examination of mining properties in the Salmon River region of north-central Idaho has disclosed a thick section of metamorphosed sediments in which sedimentary gneisses are important. These gneisses resemble underlying gneisses of probable igneous origin but are distinct from them. The dominant structural feature of the area studied is a great anticline trending northeast-southwest with large-scale thrust faulting paralleling this fold on its northwest limb. The metasedimentary series forms the northwest limb of this regional anticline. The sedimentary gneisses were apparently derived from conglomerates and coarse sands formed near a shoreline and were in part probably terrestrial. An eolian origin is suggested for some of these sandy phases. It is suggested that these gneisses and the associated schists and quartzites represent the basal part of the Upper Proterozoic (Belt series) which has widespread exposures in adjacent parts of Montana and northern Idaho. The sediments were derived from a granitic land mass lying to the north and northwest.

## INTRODUCTION

Examination of mining properties in the Salmon River region of north-central Idaho during the past year has disclosed some interesting geologic features. Of special interest are gneisses showing an apparent sedimentary origin that form widespread exposures in this area. This paper gives some preliminary notes on the stratigraphic and structural relationships of these gneisses.

The area studied is one of several square miles in extent in the vicinity of Shoup, a small mining camp located 42 miles northwest of the town of Salmon in Lemhi County, Idaho. Brief excursions were made into many parts of the surrounding territory. The location of the area is shown on the index map (Fig. 1).

## PREVIOUS GEOLOGIC WORK

During the summer of 1910 J. B. Umpleby made a general study of the mineral deposits and geology of Lemhi County, including a brief visit into the district concerned in this paper.[1] Other geologists have visited the region from time to time but have published no

[1] "Geology and Ore Deposits of Lemhi County, Idaho," *U.S. Geol. Surv., Bull. 528* (1913).

details of the stratigraphic, structural, and igneous rock geology.[2] A recent expedition of the National Geographic Society down the length of the Salmon River (October, 1935) provides many new and valuable facts about the geology of the region.[3]

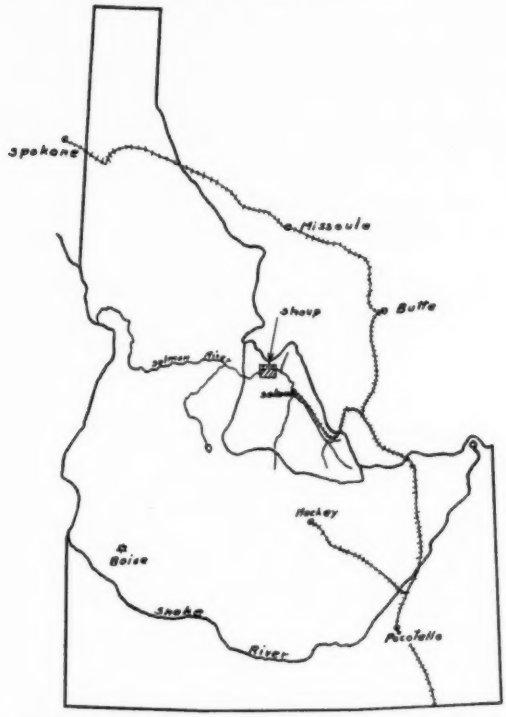


FIG. 1.—Index map showing location of area discussed in this paper

[2] George H. Stone, "An Extinct Glacier of the Salmon River Range," *Amer. Geol.*, Vol. XI (1893), pp. 406–9; George H. Eldridge, "A Geological Reconnaissance across Idaho," *U.S. Geol. Surv.* part II, *16th. Ann. Rept.*, pp. 211–76; Robert Bell, "An Outline of Idaho Geology and of the Principal Ore Deposits of Lemhi and Custer Counties, Idaho," *Proc. Intl. Min. Cong., 4th Sess.* (1901), pp. 64–80; Waldemar Lindgren, "A Geological Reconnaissance across the Bitteroot Range and Clearwater Mountains in Montana and Idaho," *U.S. Geol. Surv., Prof. Paper 27* (1904). This paper gives a few notes on the geology and mineralization of the Shoup district.

[3] Philip J. Sheron and John C. Reed, "Down Idaho's River of No Return," *Natl. Geog. Mag.* Vol. LXX, No. 1 (1936), pp. 95–136.

### TOPOGRAPHY OF THE AREA

The Salmon River valley and its numerous tributaries in the vicinity of Shoup are deep and narrow, owing to recent rejuvenation and oversteepening. The resulting steep mountain slopes bordering the river gradually pass into the more rounded and mature topography of the adjacent divides. The higher ridges and divides form remnants of an old peneplaned surface that varies in elevation from 6,000 to 8,000 feet or more. The approximate elevation of the valley bottoms in the vicinity of Shoup is 3,300 feet. The Salmon River in this region is bordered on the north and northeast by the Bitteroot Mountains, which form the divide between Montana and Idaho. The Salmon River Mountains border the stream on the south and southwest.

The recent rapid cutting by the Salmon River and the lower parts of its tributaries affords excellent bedrock exposures so that sections can be studied through a considerable vertical distance, except where complicated by faulting.

### REGIONAL GEOLOGY

In the vicinity of Shoup and over a large part of the surrounding area, especially to the west, the exposed formations are dominantly schists and gneisses, of which a considerable portion are of apparent sedimentary origin. Underlying granitic gneisses form the lower canyon walls at Shoup and for some miles up and down the river. Structurally, these lower gneisses show a very close relationship to the overlying metamorphosed sedimentary section in the strike and dip of their foliation.

Granitic stocks exposed a few miles east, south, and north of Shoup, are probably projecting knobs on a batholithic mass which underlies the area at relatively shallow depths as indicated by the multitude of pegmatitic dikes and veinlets which form a haphazard complex intrusive pattern in the metamorphic series exposed in the canyon bottoms. The exposed granitic masses are dominantly normal biotite-granite characterized by large-scale porphyrytic texture on their borders. Monzonitic phases are developed to the north of the area. South of the area especially the exposed granitic masses show incipient gneissic texture suggesting fairly intense regional

metamorphism since their intrusion. The mineral deposits of the region are roughly arranged around the borders of these intrusions and mainly in the intruded rocks.

Dikes ranging from basalt to very acid rhyolites are common over the area. Dikes of rhyolite-porphyry, diorite, and quartz-diorite-porphyry are most abundant, especially in the vicinity of Shoup, where those of the first-named type appear to be related to large-scale thrust faulting.

Over this area the metamorphosed sediments dip in general to the northwest and west and apparently form the northwest portion of a great regional anticline which contains many subordinate folds. Observations to date suggest that the granitic masses are exposed in the axis of this regional fold, although this point is by no means established.

Faults are common and locally abundant. The dominant faults are high- to low-angle thrusts forming a shear zone several miles wide. Individual faults in this general shear zone have been traced for a distance of several miles. These thrusts form the outstanding structural feature of the region and have influenced the course of the Salmon River for several miles in the vicinity of Shoup. The general strike of the fault planes is northeast-southwest, with dips of 45°–60° to the southeast. The thrust planes appear to lie on the northwest limb or close to the axis of the large regional anticline already described, roughly paralleling the trend of the fold in their strike.

The faults are marked by intense brecciation, many minor planes of slipping, and much slickensiding. Locally, the brecciated zones exceed 30 feet in width. In numerous areas of exposure the brecciated material has been recemented into a dense, hard, siliceous mass so resistant to erosion that the fault planes now form the slope of the hillside, blanketing the surface over several hundred, or even several thousand, square feet. The control of the topography by this resistant, silicified fault-plane material is a characteristic feature in the vicinity of Shoup (Fig. 2).

High-angle faults, a few of them definite thrusts but most of them normal, intersect and displace the aforementioned faults. They are of small throw in so far as observed. Many appear to be closely re-

lated to the prominent and widespread jointing of the region, and clearly represent movements along this jointing. These faults tend to intersect the regional thrust faults at acute or right angles.

The age of the faulting is important in helping to date some of the metamorphism. According to Umpleby, the granitic intrusions occurred in late Cretaceous or early Eocene time.[4] Although the thrusting movements occurred largely before the appearance of the granite, the period of intrusion followed very closely and was con-

FIG. 2.—Showing brecciated and silicified fault plane of low-angle thrust fault blanketing hill slope due to the greater resistance of the fault-plane material to erosion. Looking eastward.

temporaneous with the latest stages of thrust faulting. This is indicated by exposures studied in the Pine Creek and Big Creek districts south and southwest of Shoup, where the porphyrytic granite shows a crude but definite banding parallel to the thrust fault planes which, locally at least, pass into and involve the granitic mass. This gneissic texture was possibly formed by movements along the fault planes while the granite was still in a somewhat plastic state and is of a much younger age geologically than that of the sedimentary series next described.

## DESCRIPTION OF THE SECTION IN THE VICINITY OF SHOUP

A 1,300-foot section of the metamorphic series exposed on the north wall of the Salmon River canyon was measured and studied in

[4] *Op. cit.*, pp. 42–43.

detail just west of Shoup. This section includes 450 feet of gneiss at the base that erodes into nearly perpendicular cliffs and crags above the canyon bottom, these rocks passing upward into a series of schists, quartzites, and interbedded gneisses of sedimentary origin. Excellent exposures of all the contacts occur.

Umpleby describes the contact of the schist-quartzite-gneiss series with the basal gneiss as follows:

> The base of the Algonkian as exposed near Shoup is a fine-grained micaceous schist, locally including abundant fragments of the underlying gneiss. It gives way in a short stratigraphic distance to quartzite beds which alternate with those of schist and slate for 200 or 300 feet.[5]

He gives the following detailed description of the basal gneiss:

> Coarse-textured granite gneiss occupies a broad area in the northwestern part of the county. It is typically developed in the canyon sides about Shoup. . . . .
>
> The formation is separated from the overlying schists by a profound erosional unconformity, clearly shown in a few small exposures along the road from Shoup to the Kentuck mine. In most places weathered particles of gneiss enter into the lower few feet of the schist, forming an arkose, but elsewhere coarse undecomposed feldspar crystals appear at the very contact. The gneiss is thought to be an intrusive rock, although the local evidence of this is little more than suggestive. In a few places within the gneiss are small masses (up to 4 feet) of fine grained micaceous material. As seen in the field these areas suggest inclusions of mica schist, and such they may be, but if so they have assumed a mineralogic make-up very similar to that of the gneiss, save for a higher percentage of biotite. From the broad theories of geology, however, and as in most places where similar granitic rocks have been thought to be portions of an original earth crust they have proved to be intrusive into older sediments, the rock is believed to be intrusive. Because of its relation to the overlying Algonkian it is assigned to the Archean.
>
> The gneiss presents a striking appearance in natural exposures. Numerous large feldspar crystals, crushed into thick lenses, interleave one with another, separated only by bands of biotite which bend about them. Quartz, all of which is crushed, is scattered through the biotite bands and among fragments of feldspar. Magnetite crystals are scattered through the finer-grained parts. The feldspar crystals on account of their length, which probably averages 1 to 2 inches and over large areas 3 to 4 inches, are most resistent to weathering and present a nodular aspect on exposed surfaces.
>
> Microscopically examined, the rock is seen to be made up of orthoclase with micropegmatite, quartz, and biotite in subordinate though conspicuous amounts.

[5] *Ibid.*, p. 31.

Oligoclase, muscovite, apatite, magnetite, and diopside are accessory and decreasingly important in the order named. Many of the feldspars are crushed; many of the biotites are broken and bent and others recrystallized; quartz is fractured; and apatite crystals show many cross cracks.[6]

Umpleby's very adequate description of the basal gneiss in the foregoing section was found to apply to these rocks wherever exposed. It is important to note again that the gneissic banding in these lowest rocks parallels the dip and strike of the schistosity and bedding of the overlying formations very closely in all the observed outcrops. This relationship is too persistent and widespread not to have some meaning. These gneisses have been penetrated over a considerable area in the underground workings of the Gold Hill mine just west of Shoup, and everywhere the banding shows the same persistent moderate northwest dip and northeast-southwest strike. If the basal gneisses are of igneous origin, which is strongly suggested by their mineralogic constitution, it is probable that they were involved in the same metamorphic processes that changed the overlying sedimentary series into schist, quartzite, and gneiss.

Several sections of the formations above the basal gneiss just described were measured and studied. The section exposed on the hillslope above the high cliffs just west of Shoup is characteristic. It shows the general sequence of formations given in Table 1.

The contact of the early pre-Cambrian gneiss with the overlying quartzites is generally sharp and marked by a relatively smooth and level erosional surface. This suggests profound erosion and even peneplanation of the granitic land mass before the overlying sediments were deposited. The basal quartzites are characterized by irregular layers and lenses of coarse quartzite, locally conglomeratic, grading into schist phases. The schists frequently show much biotite, although muscovite is generally dominant. Included fragments of the basal gneiss are common. Occasional thin lenses of hornblende-schist occur.

The quartzites in the section form conspicuous, light-colored exposures above the highway on the north wall of the canyon (Fig. 3). They are dominantly light gray or white in color and generally fine grained, often showing quartz-schist phases. The bedding is gen-

[6] *Ibid.*, pp. 41–42.

erally thin, a few inches to a foot or more, and laminations are abundant. Small-scale cross-bedding is common. These quartzites grade rapidly into lenses of gneiss and muscovite schist, horizontally as well as vertically.

The gneissic lenses in the lower and middle parts of the section are composed of drawn-out pebbles and grains of quartz, often much fractured and frequently containing inclusions of orthoclase feldspar.

TABLE 1

STRATIGRAPHIC SEQUENCE WEST OF SHOUP

TOP ERODED

| | Feet |
|---|---|
| 1. Gneiss, sedimentary, with thin layers and lenses of quartzite, quartz-schist, and mica-schist. Gneiss very sandy and weathers into massive prominent local bluffs. | 500 |
| 2. Quartzites, with lenses and beds of gneiss, mica-schist, and quartz-schist. Several prominent schist beds..... | 35 |
| 3. Schists, mainly, with lenses and thin beds of white quartzite and lenses of gneiss..... | 100 |
| 4. Quartzites, with lenses and thin beds of mica-schist and gneiss. Much schist at the base of the section. Locally much biotite-schist, especially at the base of the section. Quartzites form conspicuous outcrops of light color.... | 220 |
| GREAT UNCONFORMITY | |
| 5. Basal gneiss of early pre-Cambrian age. Probably of granitic origin..... | 450+ |

Some of the nodules are composed chiefly of feldspar. Sheaves of muscovite, less commonly a combination of muscovite and biotite, are wrapped around these nodules and drawn out into long parallel stringers. The gneissic banding thus formed is parallel with the cleavage of the associated schists. The bedding of the quartzites and schists in all places noted parallels the cleavage of the schists and related gneisses.

Mineralogically these gneissic lenses and beds are characterized by an abundance of quartz in the form of nodules, grains, with a sandy matrix, the latter feature predominating in the upper part of the section. Feldspars and basic minerals are subordinate quantitatively, but locally abundant and dominant. Flakes and minute scattered fragments of muscovite are everywhere common. Especially

toward the upper part of the section, the gneisses are characterized by lamination and cross-bedding emphasized by thin layers of biotite, magnetite, and subordinate hornblende. The quartz grains and fragments are angular or, somewhat more rarely, rounded. Rounding seems more common in the finer-grained phases and in the associated quartzites. The arkosic nature of the section is everywhere characteristic.

FIG. 3.—White quartzites exposed in the cliffs above Shoup and interbedded with sedimentary gneiss and schists.

The massive-bedded gneiss in the upper part of the section is of special interest. A sandy gneiss, similar in textural features to the lower lenses of gneiss, predominates. Thin layers of schist and lenses of quartz schist and quartzite are scattered throughout this section but rarely exceed 2 or 3 feet in thickness. The outcrops of these massive sandy beds strikingly resemble exposures of Tertiary sandstone. Most of the weathered outcrops are sugary and friable, and it is difficult to realize that these formations are of pre-Cambrian age (Fig. 4). Little evidence was found to indicate that cementing material between the sand grains had been dissolved by weathering processes. The clean, washed nodules, fragments, and grains of white quartz are typical of this phase of the section everywhere. Although metamorphic processes have largely destroyed the original character of these sediments, there is still evidence of cross-bedding, some of it on a large scale. The possibility of dune sands and an eolian origin for a part of these gneisses merits consideration. Larger exposures viewed as a whole show abrupt lensing, thinning, and thickening in beds of varying texture. Where much shattered adjacent to faulting,

these gneisses weather to a brown color by oxidation along the many fracture planes.

### TENTATIVE INTERPRETATION OF THE SECTION

The section described is thought to be part of a thick series of metamorphosed sediments derived from a granitic land mass and formed at or near the shoreline of an invading sea. The abundance of minerals of igneous origin indicates the granitic source of the sedimentary material. Short transportation is suggested by the unaltered character of the feldspars and the general coarseness of the fragments. The rocks are nearly all arkosic, though locally they form a typical graywacke.

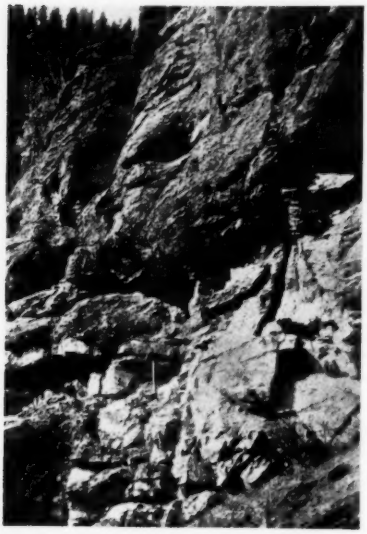

FIG. 4.—Characteristic outcrop of sedimentary gneiss exposed on the mountain slope above Shoup. Note the contact of the white quartzite with the overlying gneiss. The dip is at a moderate angle to the northwest.

Conglomerates and coarse sandstones probably formed these gneisses, as indicated by the abundance of quartz pebbles and coarse fragments. This texture, together with the cross-bedding, lenticularity, and abrupt gradation of one size of material into another, suggests shoreline or nearshore deposition. When viewed from a distance, eolian cross-bedding is strongly suggested in massive exposures in the uppermost part of the section.

The quartzites and schists were probably derived from finer sediments, shales and sands, which had been deposited in quieter waters or farther offshore. Conditions of deposition varied frequently as evidenced by the abrupt lensing and gradation of one type of rock into another. Current action is indicated by the abundance of small-scale cross-bedding.

Regional studies show that the quartzites and other finer sediments thin out to the north and northwest of Shoup and thicken to the south, east, and northeast. This suggests a land mass to the north and northwest. A few miles east and south of Shoup, the described section passes into a thick series of quartzites that are fine grained and uniform through hundreds of feet of thickness. Slates and argillites are less common in the section but locally predominant. These beds suggest more uniform conditions of deposition than in the areas to the west and north in the vicinity of Shoup. The gradation of the quartzites and schists into the massive, coarse, sandy gneisses, as the formations are traced north and northwest from Shoup, is suggestive of an approach to shoreline conditions in this direction.

Reconnaissance trips down the Salmon River canyon west of Shoup for some 30 miles showed 3,000 feet or more of bedded schists forming the canyon walls. Interbedded quartzites and gneiss are subordinate. This series was traced eastward into the section studied at Shoup and has also been traced into the quartzite-slate series of the upper Salmon River area. There is thus increasing metamorphism of the sedimentary series westward and northwestward. The great Idaho batholith, which is widely exposed in the Clearwater and Bitteroot mountains to the north and west of the area, may have been a factor in this metamorphism, but the writer is inclined to believe that much of the regional metamorphism occurred before the intrusion of this batholithic mass. This seems true at least in the vicinity of Shoup where the relationships of the granite stocks to the intruded rocks have been carefully studied.

In conclusion it is believed that the section of metamorphosed sedimentary formations exposed in the Shoup area represent the basal beds of the Upper Proterozoic (Belt series), which is widely developed to the north in Montana and northern Idaho. It is suggested that the section studied in the Salmon River region was formed close to a land mass to the north and northwest and may be in part terrestrial. The invading sea lay mainly to the south, southeast, and northeast in this area. It is believed that the section described in this paper represents one of the very few areas of the northwestern United States that shows the contact of the more ancient land mass with the Belt sedimentary series.

# CONCRETIONS IN THE FAYETTEVILLE SHALE[1]

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## ABSTRACT

In northwestern Arkansas the Fayetteville shale, Middle Chester in age, consists largely of fissile, carbonaceous clay shale with limestone lentils restricted essentially to the lower part of the formation. The shale contains argillaceous, calcareous, limonitic, and gypsiferous concretions. The argillaceous concretions are small epigenetic nodules, and occur sporadically throughout the shale. The calcareous concretions are clay-ironstone, usually septaria, and have originated by replacement of the limestone lentils in the lower part of the shale. They are the most numerous and largest concretions in the Fayetteville shale. The limonite and gypsum concretions are restricted to the lower part of the shale and represent alteration of the clay-ironstone concretions. The limonite concretions are relatively common, but the gypsum concretions are rare.

## INTRODUCTION

The Fayetteville shale, Upper Mississippian (Middle Chester) in age and ranging from 10 to 400 feet in thickness, is widely distributed in northwestern Arkansas and northeastern Oklahoma. The formation in northwestern Arkansas consists largely of black, fissile, highly carbonaceous shale. Locally calcareous clay shale comprises much of the sequence. A sandstone member, the Wedington sandstone, is generally present in the upper part of the formation. In the lower part of the formation, and very rarely in its upper part, limestone lentils are present. The shale is overlain by the calcareous beds of the Pitkin limestone (Upper Chester) and Hale formation (early Pennsylvanian), and underlain by the calcareous Batesville sandstone (Lower Chester) and Boone limestone (Middle Mississippian).

The Fayetteville shale contains pyrite as crystals and grains, within the individual beds of shale, and as nodules, and thin seams of pyrite between the shale beds.

## OCCURRENCE OF THE CONCRETIONS

Argillaceous, calcareous, limonitic, and gypsiferous concretions occur in the Fayetteville shale. The argillaceous concretions are sporadic throughout the entire thickness of the formation except in

[1] Published as Research Paper No. 446, "Journal Series," University of Arkansas.

the Wedington sandstone member. The calcareous concretions are localized in the lower part of the formation where they may occur as isolated nodules or as continuous horizons projecting as ledges from erosion slopes of the shale. The limonitic and gypsiferous concretions are likewise restricted essentially to the lower part of the shale. The gypsiferous concretions are relatively rare. Of the four kinds the calcareous concretions are the largest and most numerous.

### ARGILLACEOUS CONCRETIONS

The argillaceous concretions are normally lenticular with longest dimensions rarely exceeding 3 inches. They are brown to yellowish brown in color and consist chiefly of clay with a small amount of quartz and hydrated oxides of iron. They lie along the bedding planes of the shale with longest dimensions paralleling the bedding. The shale is usually flexed above and below the concretions indicative of their epigenetic origin.

### CALCAREOUS CONCRETIONS

The calcareous concretions appear in fresh exposure of the shale as globular, spindle-shaped, and lenticular bodies, reddish brown, brown, dark gray, or black in color, sometimes fossiliferous, ranging from a few inches to more than a foot in thickness, and from a few inches to several feet in breadth. Since formation they have almost invariably been altered to septaria with veins of gray or white calcite, which may be darkened by organic matter, oxides of iron, or iron carbonate. After exposure to weathering the calcite veins become partly or wholly altered to limonite.

Analyses of unaltered fragments and chips from unweathered surfaces of the clay-ironstone concretions indicate an abundance of carbon dioxide, iron, calcium, aluminum, and the sulphate radical. The composition of the concretions is interpreted, therefore, as a mixture of secondary calcite and siderite with clay, and calcium sulphate. They are therefore appropriately designated "clay-ironstones."[2] Seams and stringers of pyrite ramify through the interior of many of the specimens. Organic matter is invariably present,

[2] A. H. Purdue and H. D. Miser, "Eureka Springs—Harrison Folio," *U.S. Geol. Surv. Folio 202* (1916), p. 13.

often in sufficient abundance to impart a dark color to the concretions. When the concretions are broken, a fetid or petroliferous odor is often discernible. The interiors of many of the concretions are uniform in color, texture, and composition; others possess a core consisting of a mixture of iron and calcium carbonates and calcium sulphate, surrounded by a darker layer of almost pure siderite. After long exposure the concretions exhibit an outer crust of limonite.

That the clay-ironstone concretions and septaria have resulted from the replacement of beds and lenses of limestone, more rarely beds of calcareous shale, is demonstrated by their occurrence at certain and definite horizons, the juxtaposition of the concretions laterally, their orientation with longer axes parallel to the bedding, the general absence of nuclei in the center of the concretions, the continuation of the layers of concretions into unaltered limestone beds, and their restriction chiefly to the lower part of the formation to which calcareous beds and lenses are essentially restricted.

It is apparent from the composition of the concretions that the original calcareous beds have been largely replaced by ferrous carbonate and calcium sulphate. The alteration is attributed mainly to the action of percolating underground water carrying ferrous sulphate and sulphuric acid in solution derived from the oxidation of pyrite present in the shale (eq. 1).[3]

$$2FeS_2 + 7O_2 + 2H_2O = 2FeSO_4 + 2H_2SO_4 \qquad (1)$$

The oxidation takes place in the zone of weathering, and the iron-bearing solutions are carried downward by descending groundwater and brought into contact with the calcareous beds, replacing the calcium carbonate with ferrous carbonate and hydrous calcium sulphate (gypsum) according to equation 2. The sulphuric acid present reacts with the calcium carbonate to form additional gypsum as indicated in equation 3. The replacement results in an increase in volume which appears in the displacement and flexing of the shales above and below the concre-

[3] V. H. Gottschalk and H. A. Buehler, "Oxidation of Sulphides," *Econ. Geol.*, Vol. VII (1912), pp. 15–34.

tionary masses. That the process results in the formation of considerable secondary calcite is indicated by the vigorous reaction of the clay-ironstone concretions with cold, dilute acid.

$$FeSO_4 + CaCO_3 + 2H_2O = FeCO_3 + CaSO_4 \cdot 2H_2O \quad (2)$$

$$H_2SO_4 + CaCO_3 + H_2O = CaSO_4 \cdot 2H_2O + CO_2 \quad (3)$$

Although the replacement of calcareous strata by ferrous carbonate, calcium sulphate, and calcium carbonate appears sufficiently explained by the process just described, other reactions are possible, and it is by no means certain that they have not played a part in the alteration of the calcareous beds of the Fayetteville shale.

Under normal environmental conditions ferrous sulphate slowly oxidizes in acid solutions to ferric sulphate, or to a mixture of ferric sulphate and ferric hydroxide, according to equations 4 and 5. In dilute solutions ferric sulphate hydrolyzes to a basic ferric sulphate (eq. 6) which is rather unstable, breaking down to form ferric hydroxide and sulphuric acid (eq. 7). The reactions are as follows:

$$4FeSO_4 + O_2 + 2H_2SO_4 = 2Fe_2(SO_4)_3 + 2H_2O\,, \quad (4)$$

$$12FeSO_4 + 3O_2 + 6H_2O = 4Fe_2(SO_4)_3 + 4Fe(OH)_3\,, \quad (5)$$

$$Fe_2(SO_4)_3 + H_2O = Fe_2(SO_4)_2O + H_2SO_4\,, \quad (6)$$

$$Fe_2(SO_4)_2O + 5H_2O = 2H_2SO_4 + 2Fe(OH)_3\,. \quad (7)$$

But since the environment is deficient in normal amounts of air and water and is reducing, owing to the high carbonaceous content of the shales, it is probable that only very minor amounts of the ferric sulphate can form. Brought into contact with limestone ferric sulphate forms ferric hydroxide and hydrous calcium sulphate according to the following reaction:

$$Fe_2(SO_4)_3 + 3CaCO_3 + 9H_2O = 3(CaSO_4 \cdot 2H_2O) + 2Fe(OH)_3 + 3CO_2\,. \quad (8)$$

Since limonite is essentially absent from the unweathered concretionary masses, the preceding reactions are of very subordinate importance in the alteration of the calcareous beds.

It is possible that a third set of reactions may occur, depending

on the higher position of calcium relative to iron in the electromotive series. The calcareous beds go into solution in carbonated waters as the bicarbonate (eq. 9). The calcium replaces the iron of the ferrous sulphate, when the two are brought into contact, forming the soluble ferrous bicarbonate (eq. 10) which reacts with the calcareous beds to form siderite (eq. 11).

$$CaCO_3 + H_2CO_3 = Ca(HCO_3)_2 \quad (9)$$

$$FeSO_4 + Ca(HCO_3)_2 = Fe(HCO_3)_2 + CaSO_4 \quad (10)$$

$$Fe(HCO_3)_2 + CaCO_3 = Ca(HCO_3)_2 + FeCO_3 \quad (11)$$

A fourth sequence of reactions is also possible in which the ferrous sulphate is hydrolyzed to ferrous hydroxide (eq. 12), which is carried into contact with calcium bicarbonate forming by the action of carbonated groundwater on the calcareous beds and gives rise to a mixture of iron and calcium carbonates with calcium sulphate, as indicated in equations 13 and 14. The products of the reactions replace the limestone to form the concretionary masses.

$$FeSO_4 + 2H_2O = Fe(OH)_2 + H_2SO_4 \quad (12)$$

$$Ca(HCO_3)_2 + Fe(OH)_2 = FeCO_3 + CaCO_3 + 2H_2O \quad (13)$$

$$Ca(HCO_3)_2 + H_2SO_4 = CaSO_4 + 2H_2O + 2CO_2 \quad (14)$$

It is obvious that, if reactions 12 and 13 are important in the alteration of limestone, the sulphuric acid formed must be immediately neutralized if the reactions are to be completed. The acid may be neutralized by reacting with the limestone directly, (eq. 3), or by reacting with any of the bicarbonate that may be forming (eq. 14).

All four sets of reactions may be operative in replacing the calcareous beds of the formation, but the first set is regarded as of major importance. There is ample experimental evidence that siderite may be formed by replacement where iron solutions, particularly ferrous solutions, act upon limestone.[4]

Owing to the nature of the reactions, requiring both air and water in some quantity, and the compact character of the shale it seems

[4] F. W. Clarke, "The Data of Geochemistry," *U.S. Geol. Surv. Bull. 770* (1924), pp. 581–84.

doubtful whether clay-ironstone concretions can form in the shale far from the outcrop, but the transformation takes place largely if not entirely below the zone of weathering.

### LIMONITE CONCRETIONS

The limonite concretions are small, usually less than 5 inches in largest dimensions, and have been formed by gradual alteration of small clay-ironstone concretions. They are numerous in weathered outcrops of the lower part of the shale. The concretions are composed of limonite with minor amounts of clay, and are usually porous, many consisting of concentric shells of limonite loosely held together. The porous condition is due to the normal shrinkage, estimated at about 25 per cent, which results when siderite alters to limonite, and from loss of calcium carbonate and calcium sulphate in solution. In many of the larger concretions the outer part only has been altered to limonite. The interiors of the concretions consist of clay-ironstone with or without penetrations of limonite. The formation of the limonite concretions from the clay-ironstones takes place close to the surface, with air and water abundant, and continues after exposure when incomplete.

In the formation of the limonite concretions the calcium carbonate and the calcium sulphate of the clay-ironstone concretions are removed in solution and the iron carbonate may be slowly hydrolyzed and oxidized to ferric hydrate and the latter partially dehydrated to limonite, according to equations 15 and 16.

$$4FeCO_3 + 6H_2O + O_2 = 4Fe(OH)_3 + 4CO_2 \qquad (15)$$

$$4Fe(OH)_3 + nH_2O = 2Fe_2O_3 \cdot nH_2O + 6H_2O \qquad (16)$$

Equation 15 is based on the fact that ferrous carbonate will slowly hydrolyze when bathed in water free from acids, but, since underground water normally contains carbonic acid, it is more likely that the iron carbonate, at least in greater part, goes into solution as the bicarbonate, and this oxidizes to the ferric hydroxide. The reactions are given in equations 17 and 18.

$$FeCO_3 + H_2O + CO_2 = Fe(HCO_3)_2 \qquad (17)$$

$$4Fe(HCO_3)_2 + 2H_2O + O_2 = 4Fe(OH)_3 + 8CO_2 \qquad (18)$$

The ready alteration of ferrous carbonate to limonite is well known, and slowly takes place under normal atmospheric conditions.

### GYPSUM CONCRETIONS

In 1905 Adams and Ulrich referred to the presence of gypsum in the Fayetteville shale stating that "in some localities the shale contains gypsum, which occurs as individual crystals or as a coating of calcareous concretions."[5] During the present investigation the gypsum was found to be limited almost entirely to the lower part of the formation, where it is present as concentric layers and penetrations in association with limonitic and clay-ironstone concretions, as individual grains, stringers and irregular aggregates in shale beds, as isolated crystals, thin seams, and small lenses of crystals and grains between the layers of shale, and as an efflorescence on the surfaces of outcrops of the formation. Although the manner of occurrence of the gypsum is varied, no evidence was found that any part of it was of primary origin. Nearly all the gypsum has originated by the alteration of clay-ironstone concretions.

Gypsum-bearing concretions are small, usually less than 6 inches in longest dimensions, and relatively rare. They are restricted to the lower part of the shale, and are found only in fresh exposures because of the rapid disintegration of the concretions following exposure, and the easy solubility of the gypsum.

As erosion progresses, the clay-ironstone concretions are brought into the zone of weathering where gypsum and limonite are formed by the alteration of the concretions. The gypsum occurs as the outermost layer of the concretion, or it may be enclosed by outer layers of limonite or yellow or brown ferruginous clay. Concentric layers of gypsum may alternate with layers of limonite. The center of the concretion may be unaltered, and hence clay-ironstone, or it may be penetrated by stringers, seams, and irregular aggregates of gypsum crystals. More rarely the clay-ironstone centers have been altered to limonite. A few of the concretions are merely small nodules consisting of an alternation of black shale and gypsum.

[5] G. I. Adams and E. O. Ulrich, "Fayetteville Folio," *U.S. Geol. Surv. Folio 119* (1905), p. 4.

The gypsum is beautifully crystallized, colorless, and transparent, or white and translucent in layers $\frac{1}{16}$- to $\frac{1}{2}$-inch thick. In specimens that have been near the surface for some time insufficiently protected from weathering action, the gypsum has become granular or flaky. Gypsum crystallized in the interiors of the concretions is normally red, brown, or almost black.

The origin of the gypsum is attributed in large part to the oxidation of pyrite present within the concretions and in the shales adjacent to the concretions, the reaction proceeding according to equation 1. The sulphuric acid that is formed reacts with the calcium carbonate of the concretion forming gypsum, and with the iron carbonate forming ferrous sulphate according to equations 19 and 20.

$$CaCO_3 + H_2SO_4 + H_2O = CaSO_4 \cdot 2H_2O + CO_2 \qquad (19)$$

$$FeCO_3 + H_2SO_4 = FeSO_4 + H_2O + CO_2 \qquad (20)$$

The ferrous sulphate oxidizes to ferric sulphate (eq. 4) which either forms ferric hydroxide and limonite (eqs. 5–7 and 16), or more probably reacts with the calcium carbonate of the concretion to produce more gypsum (eq. 8). Some of the gypsum may form by hydration of the calcium sulphate present as a part of the concretion. Progressing gradually inward from the exterior the gypsum replaces the concretion, forming concentric layers. During the process the part of the original calcium carbonate of the concretion escaping attack by the sulphuric acid is lost in solution (eq. 9), together with some of the calcium sulphate that is being formed. Any of the original ferrous carbonate that is not altered to ferrous sulphate is converted into bicarbonate (eq. 17) and is in part lost in solution, and in part oxidized and dehydrated to form concentric layers of limonite (eqs. 18 and 16) which alternate with the layers of gypsum in many of the concretions. Since the layers of limonite and gypsum are distinctly separated, there must be some segregation of both during their formation.

The gypsum present in the interiors of the concretions is developed in part by the reaction of sulphuric acid with the calcium carbonate of the concretion. The sulphuric acid results from the

slow oxidation of the pyrite present within the concretions. Part of the gypsum results from the hydration of calcium sulphate present.

A small part of the isolated crystals and aggregates of crystals and grains, seams, and stringers of gypsum found between beds of shale and within the individual shale beds is interpreted as forming essentially in the same way as the gypsum of the concretions. The oxidation of pyrite in the shale results in the formation of ferrous sulphate and sulphuric acid which react with calcium carbonate present in the shale producing gypsum and ferrous carbonate. The latter is largely removed in solution as the bicarbonate. The larger part of the gypsum present in the shales represents the reprecipitation of gypsum dissolved from the concretions.

Although the distribution of gypsum in the shale is very irregular and largely confined to the lower part of the formation, yet it is sufficient in amount to make an efflorescence in many places where water is issuing from beneath the surface; and the shales below high-water level along many streams have a white coating of gypsum.

## SUMMARY

The Fayetteville shale, Middle Chester in age, consists largely of carbonaceous clay shale with limestone lentils mainly in the lower part of the shale. The shale contains pyrite, and is locally calcareous.

Argillaceous, calcareous, limonitic, and gypsiferous concretions are present in the shale. The argillaceous concretions occur sporadically throughout the entire thickness of the shale. Of all the concretions the calcareous concretions, generally septaria, are the largest, most numerous, and most important. They occur in the lower part of the shale. Limonitic and gypsiferous concretions are likewise limited to the lower part of the shale. The limonite concretions are common but the gypsiferous concretions are relatively rare.

The argillaceous concretions are small, epigenetic nodules. The calcareous concretions are clay-ironstone, often several feet in breadth, and have resulted from the replacement of beds and lenses of limestone in the lower part of the shale. The alteration is attributed chiefly to the action on the limestone of underground water

carrying ferrous sulphate and sulphuric acid in solution derived from the oxidation of pyrite. The limonitic concretions are small and porous, representing the gradual alteration in the zone of weathering of small clay-ironstone concretions. The gypsum concretions are small and consist typically of layers of gypsum alternating with layers of limonite or clay. More rarely the gypsum occurs as the outermost layer of the concretion surrounding a center of clay-ironstone or clay-ironstone and limonite. The gypsum has originated chiefly as a result of the reaction of sulphuric acid with the calcium carbonate of the concretion, the sulphuric acid forming by oxidation of pyrite within the concretions and adjacent shales.

# SINKHOLES OF THE CUMBERLAND PLATEAU

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In a recent paper[1] Paris B. Stockdale has described an unusual sinkhole occurring in the sandstones near the eastern escarpment of Walden Ridge, at a considerable stratigraphic distance above the Bangor limestone. The origin of this sinkhole, known as Montlake, is believed to have been the collapse of a great thickness of shale and sandstone into a cave developed by normal solution in the Bangor limestone. Although this sinkhole is probably the most striking example of its type, it is but one of a number of similar sinkholes known to occur in the Cumberland Plateau and Walden Ridge.

A similar, though smaller, sinkhole is present in Bledsoe County, Tennessee, at the western escarpment of Walden Ridge, at the head of a small hollow, tributary to Sequatchie Valley. This sinkhole is shown on the Pikeville Special Quadrangle and is located about 4 miles east of Pikeville, in the south-central portion of the central rectangle, just west of the headwaters of Coalbank Creek. It is about 60 feet deep, with a steep, clifflike eastern wall, but with a gentler slope on the western side. Although the Sewanee conglomerate, which forms the crest of Walden Ridge and the walls of the sinkhole, dips gently eastward, the western rim of the sinkhole is slightly lower than the eastern rim, and it contains many large blocks of Sewanee conglomerate, which have irregular dips.

There is no permanent lake in this sinkhole, as is the case at Montlake, but all the drainage is definitely toward it, and a small stream entering from the east wall disappears into the bottom. No spring was found to emerge at any similar elevation along the escarpment.

Owing to the thinning of the Lower Pottsville sediments toward the west side of Walden Ridge, the stratigraphic interval above the

[1] "Montlake—an Amazing Sinkhole," *Jour. Geol.*, Vol. XLIV (1936), pp. 515–22.

Bangor limestone here is not so great as at Montlake, but it is at least 600 feet. Some limestone is also present in the Pennington formation, at about 145 feet below the plateau rim,[2] but the beds are relatively thin and shaly. Because of the great thickness of relatively insoluble sediments between the limestone and the sinkhole, and because of the large, irregularly dipping blocks of sandstone on the western rim, the origin of the sinkhole might be attributed to lateral slumping of the steep side walls of the hollow, blocking the normal drainage and forming a depression behind the slumped matter. However, the absence of either a lake in the sinkhole, or a spring emerging at a corresponding level along the escarpment, suggests that water descends along joints in the underlying formations, eventually reaching a cave in the Bangor limestone.

Two additional sinkholes of this type have been reported to the writer by Ernest L. Spain, Jr., of the Tennessee Valley Authority Geologic Division. One of these, locally known as "Hell Hole," is located near Greentree, in White County, Tennessee, on a small outlier of sandstone of the Cumberland Plateau. The interval between the sandstone and the limestone here is only about 300 feet, and the sinkhole is open from the surface down to the limestone. There is no permanent lake in this sinkhole. The other sinkhole, details of which are not at hand, is located in the eastern part of White County.

These additional cases of sinkholes in the Cumberland Plateau tend to support Stockdale's theory of origin of the Montlake sinkhole by collapse of surprisingly thick overlying rocks into a cavern in the Bangor limestone. They indicate, also, that sinkholes of this type may be rather common in the Cumberland Plateau and that further investigation may disclose others. Solution by ground water in the Bangor limestone beneath the plateau has evidently been very effective.

[2] W. A. Nelson, *Tenn. Div. Geol., Bull. 33-A* (1925), p. 75.

# REVIEWS

*East African Plateaus and Rift Valleys.* By BAILEY WILLIS. Washington: Carnegie Institution of Washington, 1936. Pp. x+358; figs. 16; pls. 72. Frontispiece, in color.

In 1929 Bailey Willis went to East Africa under the auspices of the Carnegie Institution to study the rift valleys which lie sunk in the plateaus. Plateaus, he knew, characterize Africa, which thus differs from other continents where plains and mountain chains prevail. "Why plateaus and why rift valleys in plateaus?" he soon asked himself as it became apparent that the plateaus themselves hold the clues to the origin of the rift valleys.

Following an admirably presented "historical retrospect" of rift-valley theories, Willis works toward the answer to his first query in a chapter entitled "Continental Features and Their Meaning." The habit of East Africa has been to rise. A Miocene peneplain has been lifted several thousand feet into the plateau of today. The uplift has been slightly unequal and the differences have been localized in flexures and even more definitely in faults. The faults are upthrusts and their planes are vertical or nearly so.

> It is most logical to attribute a vertical shear to a pair of vertical forces; for instance to gravity and its opposite, levity. Let gravity be considered simply as a restraining force, holding things down. Let levity act to lift. Let there be a boundary between an area within which gravity controls and an adjacent area beneath which levity is effectively greater than gravity. The two forces will then constitute a force couple, tending to shear the crust along the boundary between the areas in which they respectively predominate. Any displacement on the shearing plane will then appear as an escarpment, a fault in the surface.
>
> The above is the simplest mechanical explanation of the uplift of the plateaus of East Africa and of their warped and faulted surface. It is incomplete, of course, because no reason is given why levity should be greater than gravity within the uplifted area, but that relation appears to be a fact with which eventually to approach the deeper problem.

The rift valleys are best developed around the margins of the central Tanganyika Plateau, or central portion of the vast African uplift, where levity has been greater than gravity. This central plateau is of oval form, about 700 miles from north to south and about 500 miles from east to

west, with an altitude of 3,000–6,000 feet. A little north of the center lies Lake Victoria Nyanza in a shallow, saucer-like depression. Taking the circumference of the central plateau at approximately 3,000 miles, Willis finds, by measuring the lengths of the marginal faults, that they aggregate between one-half and two-thirds of the total. Over that proportion of the margin, the outer crust has been strained to the breaking-point by the uplifting force; elsewhere it has been flexed.

Proceeding to the study and interpretation of the specially significant structures in this marginal belt, Willis takes up in turn the Tanganyika trough on the southwest, the western Ruanda highland, Mount Ruwenzori on the northwest, the northeastern flank around Lake Rudolph, the Kenya half-oval on the east, and the Iringa-Livingstone arch on the southeast. For each of the rift valleys he considers carefully the telltale features and from them infers the processes which have been operative. These are designated the determinants and are found to be vertical uplift, vertical downdrag, horizontal compression, shearing on vertical planes owing to rotational stresses, and vulcanism.

Compression outward from the Lake Victoria area in the center of the oval will seemingly account for some of the structures along the periphery. The upwedging of the Ruwenzori block on the northwest and uplift of the Iringa-Livingstone arch on the southeast are readily explained by centripetal compression acting normal to the trends of these peripheral structures, and the north-south structures of the Kenya half-oval are consistent with it. The Tanganyika trough south-southwest of Lake Victoria is explained in part by a force couple, with southerly shove from the Victoria saucer and northerly shove, or resistance, from the Upper Congo region, which has given a strong horizontal shift along nearly parallel faults trending northwest-southeast. Horizontal compressive forces, however, play only a subordinate part in the development of the rift valleys. The causes of the vertical uplift of the plateaus and the associated downward drag are in turn the main causes of the rift valleys. The underlying condition is thought to be a deep-seated one, affecting the stability of the continent.

Willis postulates that "the crescentic downwarps and rift valleys lie in the margins of eruptive elements of the continent; that each such element represents a magma basin at a depth below the bottom of the continental layer, that is, an asthenolith; and that the rounded or oval form is the form assumed by the asthenolith as it developed its molten mass." He regards the central plateau as the cover of an active asthenolith forming from the heat of radioactive disintegration. Passage into the liquid

state involves an increase of 5 per cent in volume with vertical swelling and lifting of the cover. The lifting effect would be defined either sharply by a vertical shear or fault or more gradually in a flexure, both of which characterize the margins of the plateau. In addition, an accompanying downdraft on the margins (explained in detail with the help of a diagram) would produce a simple downwarp or, if delimited by faults, would develop a rift valley. The trough thus formed may represent either an absolute depression below the original peneplain level or may merely have lagged behind in the elevation of the plateau.

The Lake Albert trough is a simple downwarp, partly defined by upthrust normal faults, whose bottom has lagged behind, but probably is not absolutely depressed. The Lake Edward-Kivu depression is a moderately developed downwarp but very slightly faulted. It appears to have lagged, but not to have been absolutely depressed, in the vertical movements. Lake Tanganyika occupies a double trough, the two parts being offset with respect to each other. Both were absolutely depressed, the bottoms having been drawn down. The slopes of the trough and the adjacent plateau exhibit marked evidences of rotational shear and compression in the form of vertical major faults and overthrusts. This is the largest, deepest, and most typical of the East African rift valleys. The complex Gregory (or Eastern) rift valley has lagged behind the uplifted areas on either side, but there has apparently been no depression of the floor below the original position of the peneplain. Attributing the uplift of the plateau to swelling of an hypothetical asthenolith, Willis regards the subsidence of these annular segments around it as the result of failure of the walls of the molten mass. The failure of the walls is thought to be due, in turn, to intrusion of magma beneath segments outside the magma chamber and local relaxation of the support which held the walls in position.

The Nyasa and Luangwa troughs lie beyond the confines of the central asthenolith, which is bounded by the other rift valleys and must be ascribed to other asthenoliths or to entirely different conditions. It is probable that they are marginal effects of a less active or deeper seated asthenolith beneath the Rhodesian plateau, according to the hypothesis here developed.

In addition to uplifting and lag, evidence of considerable horizontal compression was observed along several of the rift valleys peripheral to the Lake Victoria basin. The direction of the pressure having been radial suggests an expansion of the disk underlying the basin. On the assumption of an underlying asthenolith, metamorphism of the rock cover is logically to be expected; and Willis believes that growing crystals, with long dimensions dominantly horizontally outward would cause the disk

to expand radially and thin vertically. An expansion of but 1 per cent in a 200-mile radius would suffice to wedge up the marginal Ruwenzori block the 2 miles which it appears to have risen. This application of Willis' theory of metamorphic orogeny is frankly treated as speculative, as, at present, theories of the deeper dynamics of the earth necessarily must be. It is in accord with the surface geology; further experimental and theoretical work presumably will eventually decide as to its validity.

Following the masterly development of the major problems of the study, their correlation in a comprehensive survey of the whole, and the unfolding of a genetic hypothesis tying everything together, in the first hundred pages, the remainder of the volume more exhaustively treats, in the light of what has gone before, the coastal region of Tanganyika, the plateaus of East Africa, the many individual rift valleys, vulcanism, earthquake conditions, and gravity tests. High artistic skill and versatility in presentation, with significance and purpose apparent in nearly every detail introduced, make interesting reading of the factual descriptions. These are models of their kind; fitted together they give the clear dynamic picture which they are intended to convey. The reader feels that the descriptions, with interpretative exposition, are essentially the argumentation for the conclusions reached by the author. There is no laboring of points.

*East African Plateaus and Rift Valleys* is truly a commanding work. Particularly admirable is the author's method of attacking the much-debated and baffling problem of rift-valley origin, by considering the larger setting and studying first the region's characteristic habits of dynamic behavior which naturally are to be expected to dominate and govern the mechanics of formation of the smaller units of structure, such as individual rift valleys, etc. These basic studies give a command of the problem as no other method of approach can. The entire investigation has been carried through with great skill and care. Many of the illustrations are superb examples of photographic art.

R. T. C.

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*Les Roches sedimentaires de France—roches carbonatées (calcaires et dolomies)*. By LUCIEN CAYEUX. Paris: Masson et Cie, 1935. 2 vols. Pp. 463; pls. 26. Fr. 200.

Professor Cayeux offers in his two quarto volumes a complete discussion of carbonate rocks, which bids fair to become a standard reference on the subject. In the first part of his work the author discusses the calcareous rocks, beginning the section with descriptions of marine limestones. Here are included a wide variety of rocks, from coral sands through vari-

ous types of chalk to limestones proper, subdivided and classified largely on the basis of chemical composition and origin, i.e., whether organic or inorganic. These largely descriptive portions include the first 135 pages of the volume. The next section discusses the structures of the calcareous rocks, ranging from crystalline and crypto-crystalline through conglomerates and breccias to nodules and oölites. The breccias, for instance, are classified into original and secondary breccias, and oölites are discussed both from the point of view of their descriptive aspects and theories of their origin. Pseudo-oölites are described, as are various "aberrant" structures, such as cone-in-cone and stylolites. The section closes with a relatively short discussion of lacustrine limestones, comprising their formation, analysis, and structures. It should be mentioned that the several aspects of the rocks discussed in the text are illustrated with excellent collotype plates at the end of Volume II.

The detailed treatment of the calcareous rocks takes up the entire first volume (some 325 pages), and the second volume devotes about 125 pages to the dolomites. The remainder of the second volume is devoted to plates. Magnesian limestones and dolomites in general are classified into types depending upon such factors as whether the dolomite is present as a cement in the limestone, whether it is present as inclusions, or whether it occurs as independent crystals. After a description of typical dolomitic rocks, there follow a section on structures and finally a lengthy review of the origin of dolomite. This last section is an excellent discussion of the subject and includes the conditions of direct dolomite formation, dolomitization subsequent to deposition, and a general summary of the subject.

The literature on the calcareous rocks is so widely scattered that sedimentary petrologists in particular and geologists in general will welcome this compilation of the outstanding characteristics of this large class of rocks. Dr. Cayeux has a lucid style of writing, the volumes are well printed, and the illustrations of thin sections are masterpieces of reproduction. Students of sedimentary rocks cannot afford to neglect this work as a fundamental reference in their studies.

W. C. KRUMBEIN

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*The Snellius Expedition*, Vol. II: *Oceanographic Results*, Part II: *Soundings and Bathymetric Charts*. By P. M. VAN RIEL. Utrecht: Kemink en Zoon, N. V., 1934. Pp. 63; figs. 31; pls. 6; charts 16.

This first section of the comprehensive report of the Snellius Expedition was published to make available the new soundings and bathymetric charts of the expedition to the eastern part of the East Indian Archipelago.

The text first concerns itself with the method of preparing the charts and the sources of supplementary information. Next follow detailed discussions of the various seas and troughs in the area studied, including information concerning potential temperature and salinity. This information is shown in the text figures, and there also are true scale profiles to show actual bottom slopes. Some idea of the added factual data gleaned by the expedition may be had from the author's report that available soundings were increased from 3,500 to 35,000 during the expedition.

The work is well illustrated, and the detail charts are printed in colors. The principal map is a bathymetric chart of the entire area on a scale of 1:2,500,000. The entire volume is well printed, and the colored charts are clear and sharp.

W. C. KRUMBEIN

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"The Physiography (Geomorphology) of Western Australia," by J. T. JUTSON, in *Western Australia Geological Survey Bulletin 95*. 2d ed., rev. Perth: Fred W. Simpson, 1934. Pp. xvi+366; figs. 147; pls. 5.

Despite its areal exent of 975,920 square miles, the state of Western Australia has but little physiographic variety; nevertheless, this report is of interest to the physiographer because of the interesting information pertaining to the arid erosion cycle. The state consists of a vast truncated upland known as the Great Plateau, which rises gradually from the narrow coastal plains. Mesozoic and Cenozoic sediments are for the most part limited to a fringe around the coast, while the remaining area is of pre-Cambrian or Paleozoic rocks, igneous rocks predominating in the south and sedimentary in the north. The drainage is in part to the sea (exoreic), in part to the interior (endoreic), and some areas have no surface drainage at all (areic). The interior drainage area, which covers all the interior of the state, is arid and practically riverless, the small creeks running into extensive shallow basins, called "salt lakes," which are dry except after unusual precipitation. The great plateau is thus in a cycle of arid erosion except around the edges, where greater rainfall causes normal erosion. There are no mountains of the folded type; the high points and ridges are either monadnocks or fault blocks.

The author divides the state into nine physiographic provinces. At the north the provinces Fitzroyland and Ordland, valley lowlands, border a great horst, the province of North Kimberley, where the upland is in the process of dissection by stream action. On the northwest and west coasts are the provinces of Pilbaraland, Murchisonia, and Swanland, which in-

clude the coastal plains and the stream-dissected margins of the Great Plateau.

The provinces of Salinaland and Sandland comprise most of the area of the state and constitute the endoreic regions, where the present arid erosion cycle is far advanced toward a new plateau, above which residuals of the old, or Great, plateau stand as mesas and buttes. Salinaland is one of the world's great gold-producing areas. The alluvial gold is found in what are called "deep leads"—old stream channels so choked by detritus as to afford little surface indication of their existence. These old lines of drainage antedate the last uplift of the region. The land forms of Salinaland include such oddities as "natural quarries" (circular breakaways), "stonefields," "desert pavements" (angular quartz fragments derived from quartz-lode outcrops and strangely arranged in "mosaic" pattern), and "gnamma holes" (large holes in granite caused by differential solution in connection with wind scouring).

Sandland, the larger of the two inland provinces, differs from Salinaland in the absence of salt playas and in the great amount of sand, differences which are traceable to the different lithology of the two regions.

Euclonia, the remaining province in the southeast corner of the state, has no surface drainage, being a limestone plateau in an early stage of an underground drainage cycle. The limestone is cavernous, with abundant large sinks called "dongas," which support a luxuriant vegetation in contrast to the more arid surroundings. Interesting also are the "blowholes," circular pipelike openings through which air often rushes strongly enough to keep a felt hat suspended on the jet. At other times there is a strong suction.

Topographic maps of Western Australia are very incomplete at present, but this situation is being remedied.

F. C. MacKnight

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"Physiographic History of the Yangtze," by G. B. Barbour, in *Memoir of the Geological Survey of China*, Series A, No. 14. Peiping, 1935. Pp. 112; figs. 24; pl. 1.

Flowing from an elevation of about 17,000 feet in Tibet across the whole of China proper to the sea, the Yangtze Chiang is the principal artery of east-west transportation through the heart of China and has opened up to study a long section of its geology. The swift Upper Yangtze drops 16,000 feet in 1,600 miles; the 600 miles of navigable Middle Yangtze are through the successive gorges of beautiful Szechuan, while for 1,000 miles

below Ichang the Lower Yangtze moves quietly as a mature stream across a region of comparatively low relief. The present study covers the 1,400 miles of Middle and Lower Yangtze from Chungking to the sea and only reviews the observations and conclusions of others regarding the complex history of the river above this point.

During the Cretaceous period central and eastern Szechuan were receiving from a recently uplifted mountain belt to the west a thick succession of red clastic sediments from which has risen the well-known name, Red Basin of Szechuan. At the close of the Cretaceous, or early in the Tertiary, the Yenchan disturbance folded western Szechuan and adjacent parts of Tibet into mountains, wrinkled the Red Basin gently, and folded eastern Szechuan into ranges of the Jura type. Later came peneplanation, and over the peneplain surface the early Yangtze flowed. The author favors the hypothesis that the present Yangtze is a superposed stream and gives reasons for doubting that the river was antecedant to the folds. It did, however, maintain its course through several subsequent episodes of uplift, faulting, and warping. The various stages have been worked out in detail and are well described and effectively discussed in this report. Particular attention has been given to the physiography of the Red Basin, drained largely by tributaries of the Yangtze.

This study is the initial one of a project, originating with the late Dr. Davidson Black, for correlating the Pliocene and Pleistocene history of the various regions of China on the basis of the land forms developed at different stages. A very definite parallelism has been found between the developments in the Yangtze and Huangho basins, and the report gives a tentative correlation table of the Cenozoic physiographic stages for various areas from the Yangtze to Shantung and Mongolia.

R. T. C.

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